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MONTHLY WEATHER REVIEW

JAMES E. CASKEY, JR., Editor

Volume 87
Number 7

JULY 1959

Closed September 15, 1959
Issued October 15, 1959

WATERSPOUT-TORNADO STRUCTURE AND BEHAVIOR AT NAGS HEAD, N.C., AUGUST 12, 1952

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[Manuscript received June 13, 1957; revised June 29, 1959]

ABSTRACT

This note describes (1) a situation leading to the anticipation of waterspout formation, (2) three spouts which subsequently formed, and (3) the passage of the third spout over the observer. An interpretation of the first spout's formation, and of the third spout's lower structure, is proposed.

1. WEATHER CONDITIONS

The formation of tornado-waterspouts was expected at Nags Head, N.C., shortly after 1500 EST, August 12, 1952, when conditions resembling those of July 8, 1942, and June 20, 1945, which then led to waterspout formation [1], were observed. The situation is shown in figure 1b; a warm NE sea breeze at Nags Head, the location of the observer, was opposed by a warm SSW wind, with a solid line of convective clouds between. Figure 1a shows the background events leading to the formation of the cloud line. The clouds moved almost from the WSW with a slow drift toward Nags Head. Above the NE breeze, fractocumulus moved from the SW, and also cumulus from the WSW. Temperature in the NE breeze was 90° F., dewpoint 78° F., and pressure was 30.00 inches and falling. The temperature of the sea surface at shore was 72° F.

2. FORMATION OF VORTICES

Vortex circulation was expected if the NE breeze reached the cloud level (fig. 2b). As rain began falling from the rear of the cloud line (cross-hatched area in fig. 2a), I took the first 35-mm. color photograph of the tornado sequence. The next 20 minutes were uneventful, until suddenly fractocumulus shreds moving from the NE, as shown in figure 2a, appeared below the nearer edge of the cloud base, opposite the rain. The cool air dome accompanying the rain (fig. 2b) could have been the mechanism which at this point lifted the NE sea breeze into the convective cloud cell where its energy contribution (horizontal momentum favoring the start of cyclonic circulation, as in fig. 2a, and latent heat release favoring vertical exhausting) appeared to trigger the dramatic events that followed.

From here on things happened fast. The shreds grew up into the cloud base, and almost immediately a tornado cyclone was born within the cloud, causing the entire width of the base opposite (that is, north of) the rain to rotate in a circle 2 or 3 miles across, the nearer edge moving from the NE and the whole disturbance accelerating to relatively high speed. The straight edge of the cloud line was quickly bent into a wave, so that the nearer edge retreated toward the SSE (figs. 2a and 3). In a few minutes a new rotation appeared along the back of the

[1] Cooperative Weather Observer, U.S. Weather Bureau.

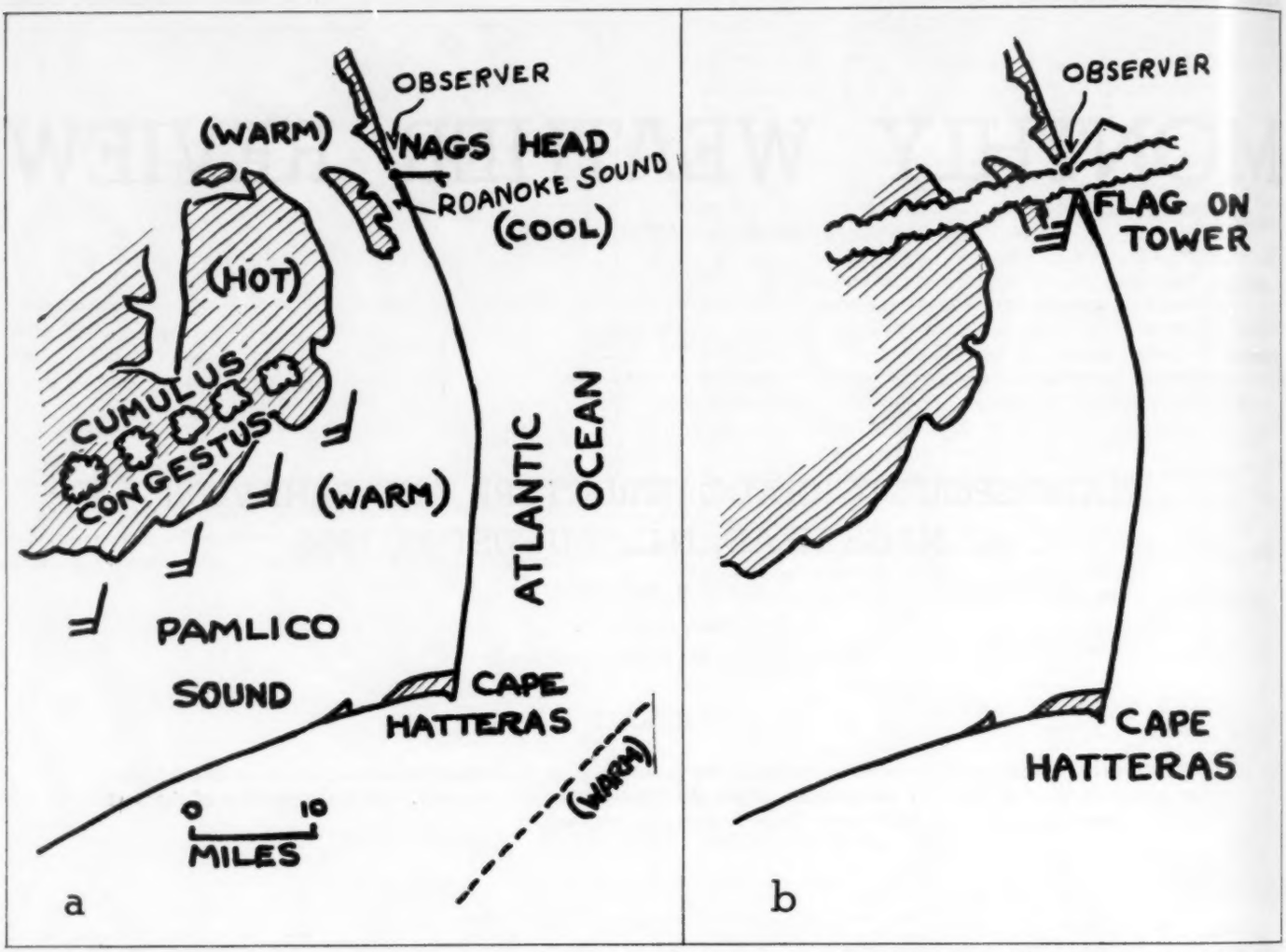


FIGURE 1.—(*a*) Distribution of surface heat and surface wind, and location of first cumulus congestus, 1230 EST, August 12, 1952. (*b*) Location of developed cumulus congestus convective line, and adjacent surface wind, 1500 EST, August 12, 1952.

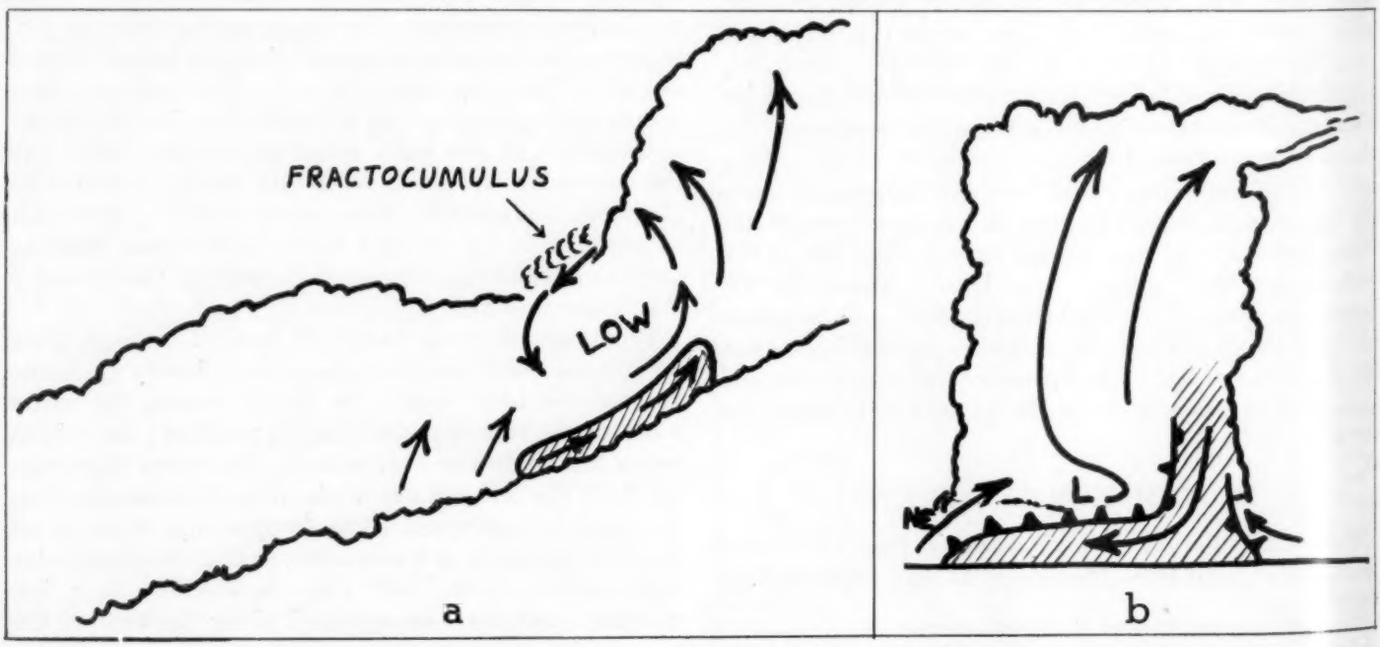


FIGURE 2.—(*a*) Plan sketch (just above cloud base level) and (*b*) elevation sketch of storm cloud line as rotary motion began.

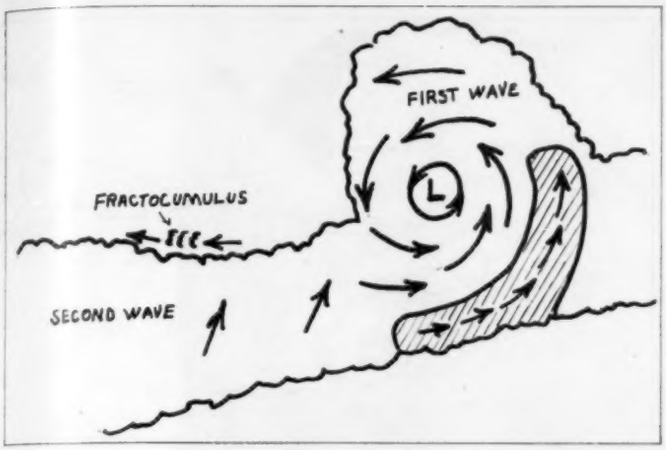


FIGURE 3.—Plan sketch of storm cloud line (just above cloud base level) a short time after figure 2.

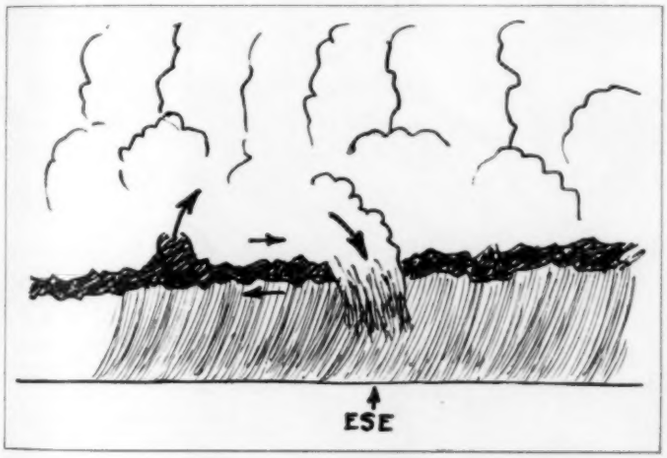


FIGURE 4.—Development of rotary movement with inclined axis in cumulonimbus base. Drawn from memory.

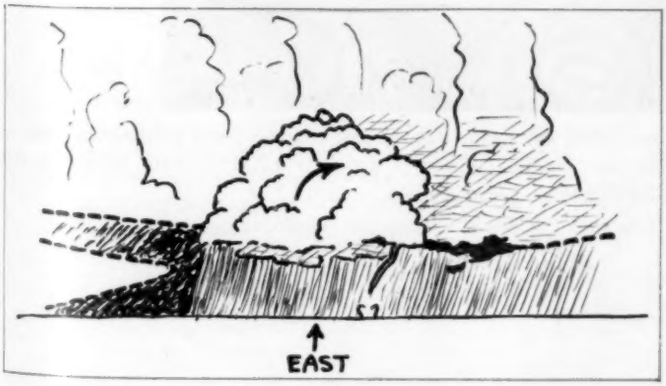


FIGURE 5.—Appearance of first waterspout depending from small rotating cumulus at rear of wave in cumulonimbus line. Drawn by the author from a 35-mm. color transparency.

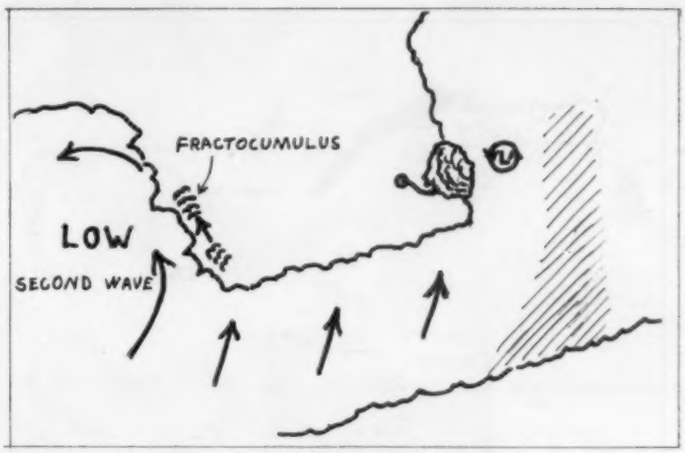


FIGURE 6.—Plan sketch of cloud line (just above cloud base level) following figure 3 and at time when first spout (right center) had formed. Rain distribution shown by cross hatching.

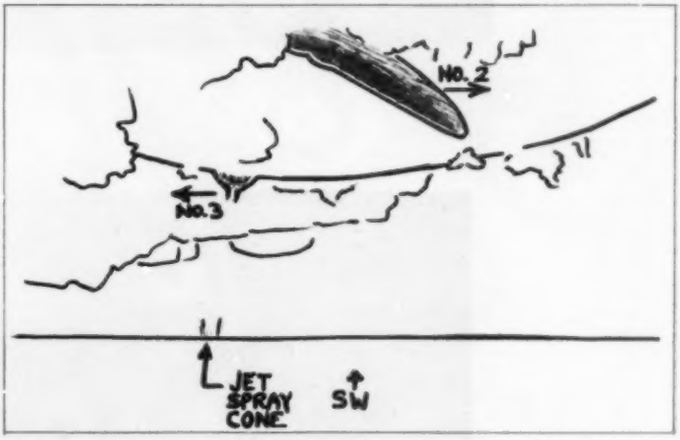


FIGURE 7.—The appearance of the second and third waterspouts depending from rotating, bowl-shaped depression at base of the second wave in cumulonimbus line. Drawn by the author from a color transparency.

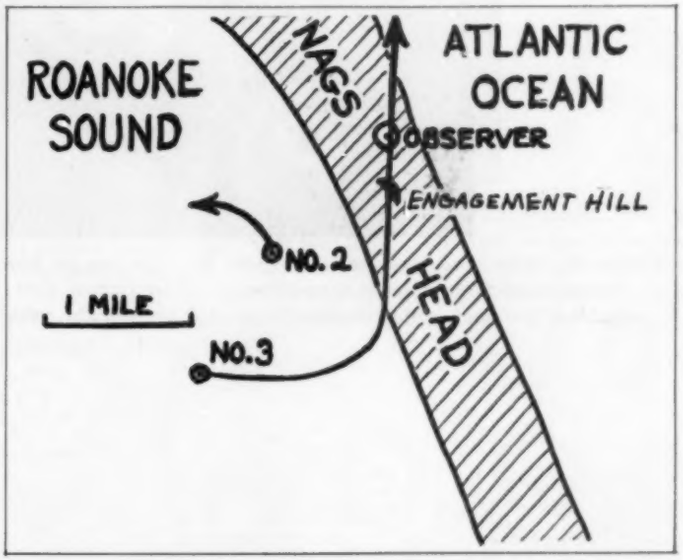


FIGURE 8.—Tracks of waterspouts 2 and 3.

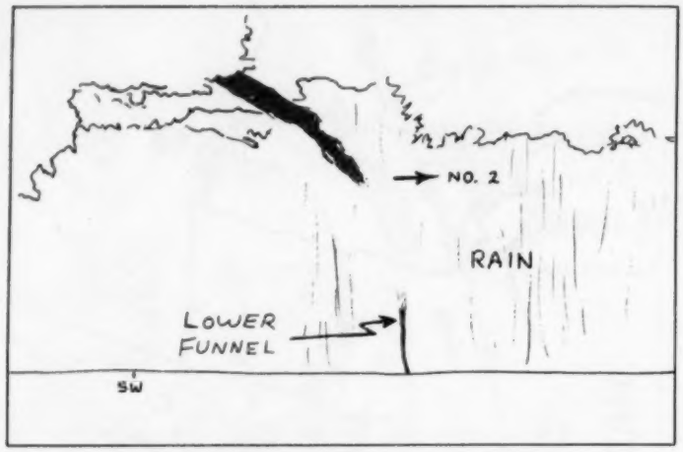


FIGURE 9.—Upper and lower vortex clouds of waterspout no. 2, which merged into a continuous vertical tube as the spout died. Rain is falling from the rear of the second wave. Drawn by the author from a color transparency.

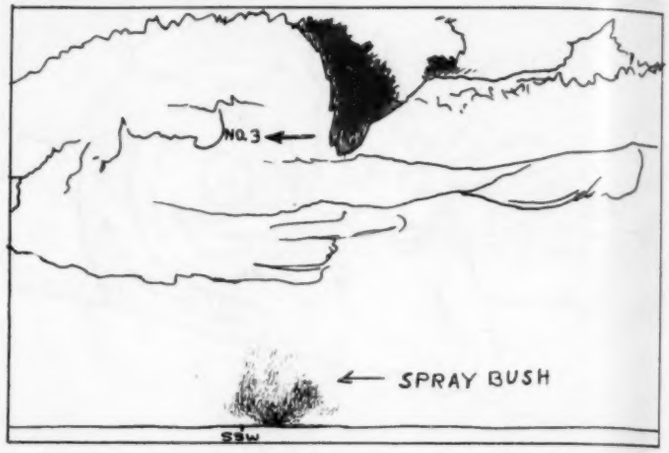


FIGURE 10.—Waterspout no. 3 moving east with thick funnel and well-developed spray bush about 400 ft high. Drawn by the author from a color transparency.

FIGURE 11.—The sand bush of waterspout no. 3 on top of Engagement Hill, 600 yards distant and moving north toward the observer. Reproduced from a color transparency. The lighter part of the sky at the right is a heavy curtain of rain following close behind the spout, and probably illuminated by sunlight reflected from water.

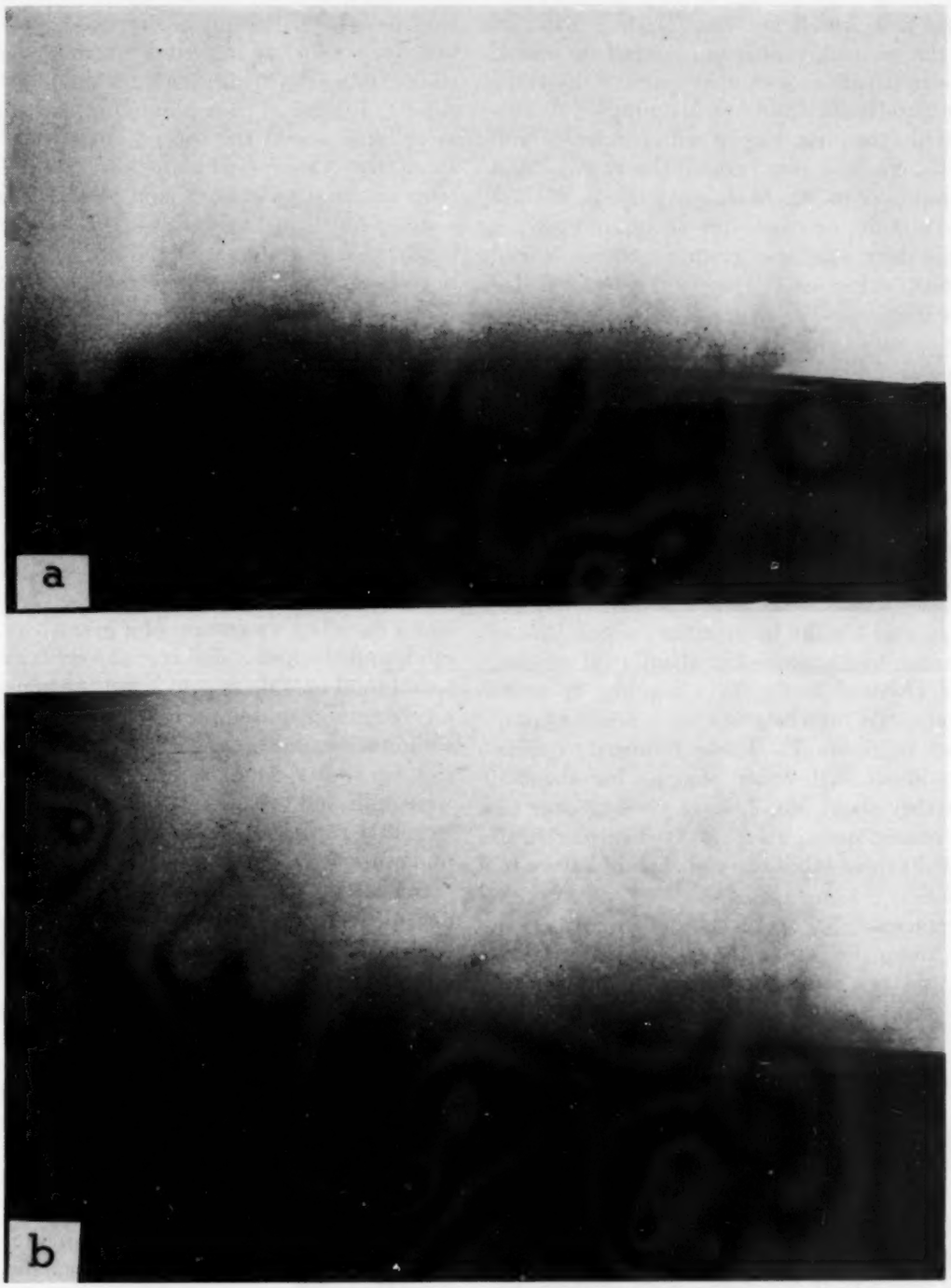


FIGURE 12.—(*a*) The sand bush of waterspout no. 3 at the base of Engagement Hill, about 400 yards distant. Reproduced from a color transparency. The heavy rain curtain seen in figure 11 has spread rapidly eastward and fills almost the entire sky area in this picture, giving the illusion of a clear sky. (*b*) This reproduction from the color shot shows the sand bush about 300 yards from the camera. The core wall reached near the top of the picture in two peaks. The outfall ring reaches to the edge at the right and nearly so to the left. The jet core shows dense at the center, between the electric wires and the black foreground. The convergence disk lightens the surface of the hill near its foot under the outfall ring.

wave, within the first whirl, but smaller, and with its axis inclined to the vertical, so that one part of the cumulonimbus edge was lifted, and another part to the right forced to plunge rapidly earthward as in figure 4. As this motion accelerated, the first larger whirl quieted, and presently from the concentrated turmoil the vortex cloud suddenly appeared over its whole length (that is, did not build downward)—thin, crooked, quivering, and silvery white against the dark rain background. Spray leaped up like a white bush at its base. The cloud tube depended from a soft-appearing, slowly turning cumulus which protruded westward and slightly downward from the lower wall of the cumulonimbus in the rear of the wave (figs. 5 and 6). The wave moved ENE and disappeared in rain, while the spout lagged behind and elongated horizontally, and finally disappeared about 10 minutes after its beginning.

Meanwhile a new wave developed to the SW of my location, with fractocumulus skirting its base, moving from the E (fig. 3), then from the SE (fig. 6) as the wave increased its activity. Soon the base of this wave was in rapid rotation, 2 or 3 miles in diameter, which *quieted* when two new tornadoes appeared at about 1630 EST over Roanoke Sound (located in fig. 1a), roughly ¾ and 2 miles, respectively, SW of the observer. Their appearance is sketched in figure 7. These followed counterclockwise orbits about each other (fig. 8) for about 10 minutes, the farther spout, no. 3, later passing over the observer. The nearer spout, no. 2, evolved into separate tubes as in figure 9; these tubes merged shortly before the spout dissolved.

In its early moments this spout showed a brief, wavelike down-rush along the upper funnel for two or three seconds, which did not lengthen the upper funnel, and which was not seen to affect the lower funnel. It accelerated from rest, and decelerated back to rest, quickly and smoothly. It was much faster than the normal rotary motion of the funnel; in fact, while it lasted, no rotary component of motion was even discernible—it looked to be entirely longitudinal with the vortex axis. Later, a similar less intense uprush was observed. It looked like an adjustment of internal pressure whose velocity along the funnel varied directly with height above the ground. The maximum velocity, if the cumulonimbus base was 2,000 feet high, was at least 500 m.p.h. With these two exceptions the spout cloud showed no evidence of vertical movement. No spray bush was discernible with this spout.

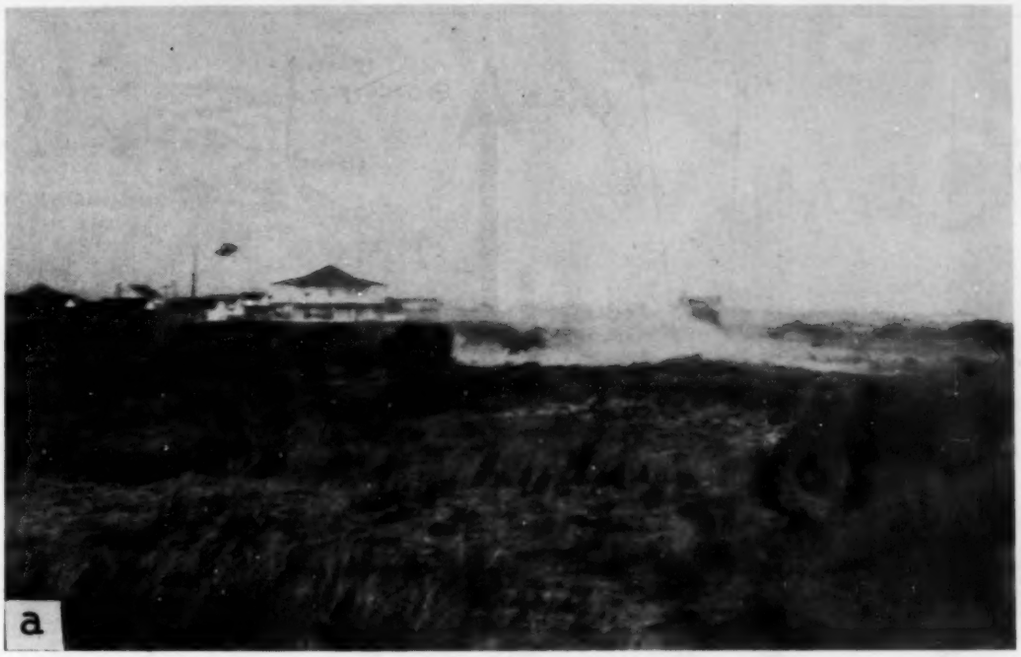
3. EVOLUTION OF TORNADO

Meanwhile spout no. 3, with a short thick funnel aloft and a spray bush about 400 feet high as in figure 10, moved eastward as shown in figure 8, and disappeared beyond Engagement Hill (located in fig. 8), a 60-foot sand dune 2,000 feet south of the observer. Its dissolution was, mistakenly, assumed. Two minutes later the top of Engagement Hill suddenly erupted into a whirling yellow monster of flying sand 200 feet high and 600 feet wide. It was bearing down on the observer at about 35 m.p.h. Figure 11 is a photograph of the whirling sand as it came across the top of Engagement Hill. It was about 600 yards distant at the time. Twenty seconds later the faint swishing sound became a heavy roar as the waterspout-turned-tornado advanced across the grassy, bushy field at the base of the hill. It was tossing thick-branched myrtle bushes aloft and reports like gunfire occurred as their roots severed. The appearance of the whirling sand is shown in figure 12a, a photograph made when the base of the tornado was about 400 yards distant. Another photo, figure 12b, shows the whirling sand bush when about 300 yards away. Details of its structure are described with the figure. Another 15 seconds and the observer's house, and probably the observer, began to shake in the wind of the tornado, and as the last photograph was made, when the tornado was only 90 yards distant as in figure 13a, brief refuge was sought indoors. In the dream-like confusion of the next several seconds there was a detached awareness of a great roar, a blur of flying sand, and the house shaking, objects tumbling inside, and a dull thud outside as a door left the house and overturned a refrigerator on the porch in departing. There was also a shower of timbers about the house, ranging in cross section up to 3 x 8 and 2 x 12 inches. Almost all of them were split and broken by the storm, yet they were lowered so gently to the earth that the soft sand around them was undisturbed.

Immediately after the surface manifestation of the tornado had passed, the short, suspended funnel was observed to be still south of the house, about 60° high, but in a few seconds it passed overhead toward the north. The surface whirl, retreating to the north, resembled a flock of circling birds with its temporarily suspended boards, boxes, garbage cans, etc. None of these objects, on falling, was identified over 450 feet from its original position. The twister passed out to sea a half-mile north of the observer, and briefly developed a cloud tube from sea to clouds, 50 to 70 feet in diameter, before disappearing entirely.

During the passage of the tornado in the vicinity of the house a 400-lb. boat, containing 1,000 lb. of rainwater had been tossed into telephone wires, then through a fence, and its parts carried 450 feet. Piles of fish boxes and lumber had been scattered about 400 feet. Eight buildings suffered minor damage. An iron pump in a fenced garden was sheared from its pipe and driven hard into the sand 10 feet away, and the surrounding fences were flattened outward. Details of objects moved and the path of the tornado are shown in figure 14.

Surface winds were light E to SE just before the storm hit at 1640 EST. Five minutes later torrential rain without hail, but with very severe lightning and thunder, was blown by a S wind of force 4, which in another five



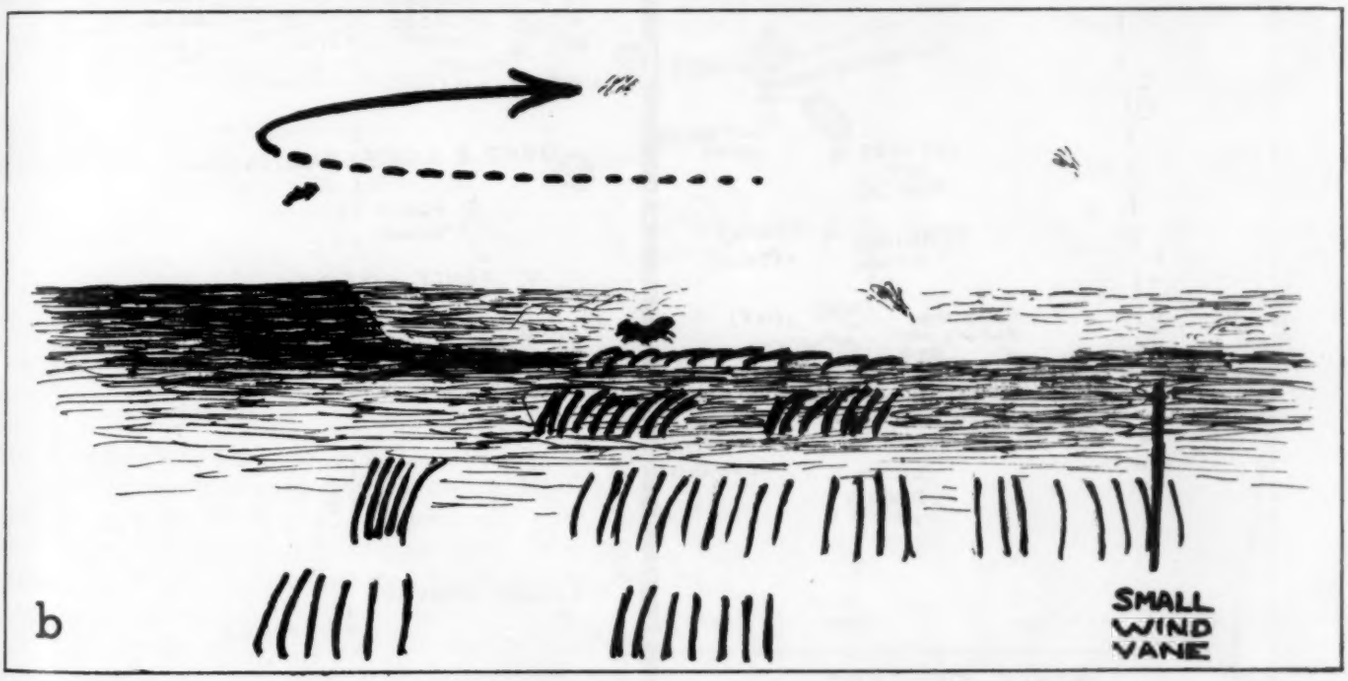


FIGURE 13.—(*a*) Reproduction from a color transparency showing view of tornado center about 90 yards distant, with observer in area of strong convergence. The tall beach grass, and wind vane turned endwise and therefore invisible, indicate how airflow was toward the center without cyclonic movement in outer part of convergence zone. Center is over shallow ditch located at bottom of figure 14. (*b*) Simplified sketch drawn from (*a*) to emphasize representative tufts of bending grass, and the wind vane support at the right.

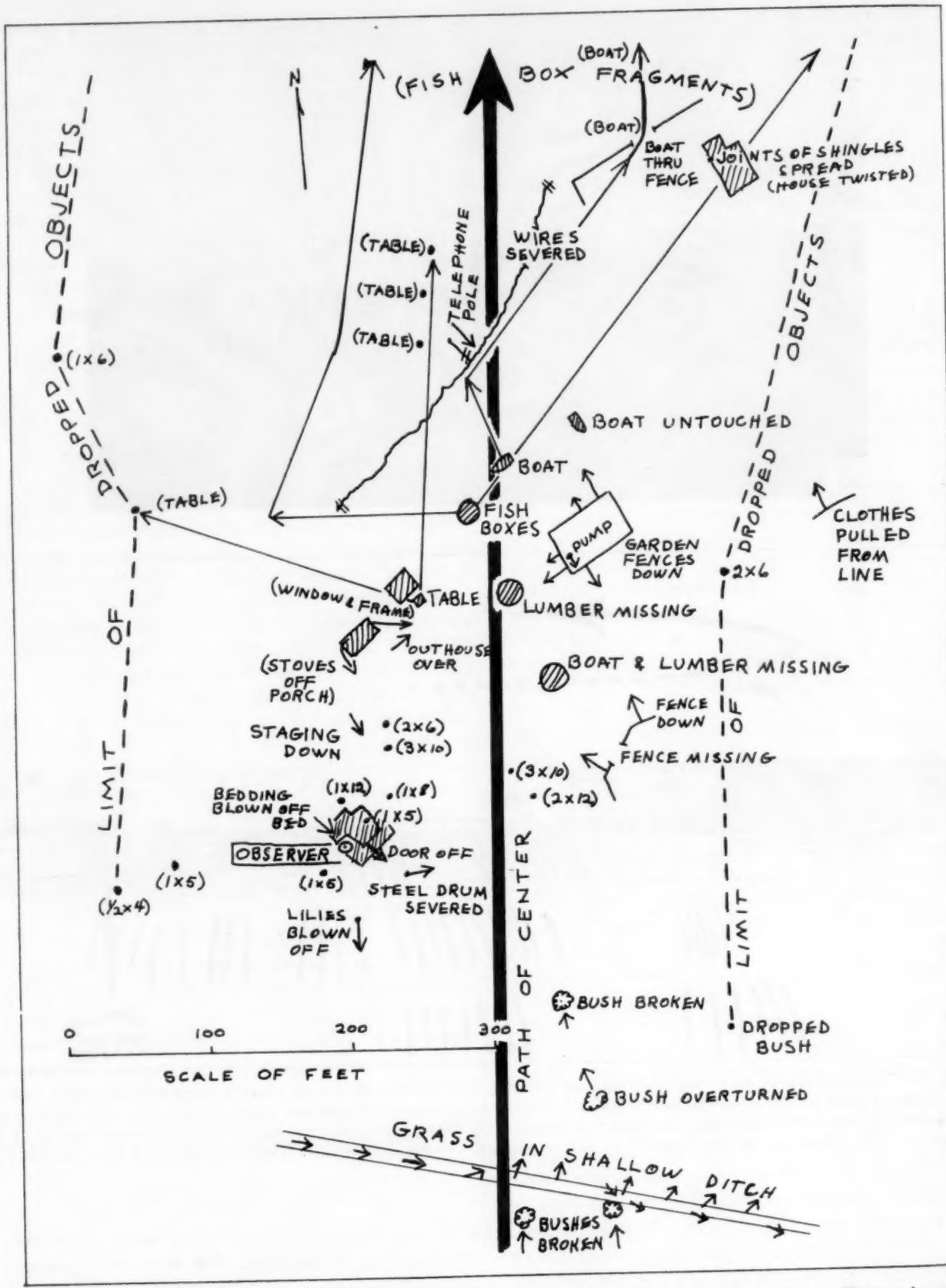


FIGURE 14.—Map of tornado's effects in observer's vicinity. Names in parentheses indicate dropped objects. Figures in parentheses show cross section dimensions in inches of dropped boards and timbers. Short arrows show direction of action. Long arrows show distribution of dropped fragments from identifiable sources.

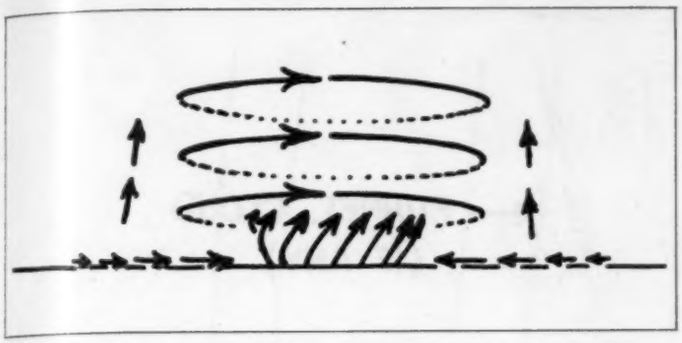


FIGURE 15.—Motions in the spray and sand bush as seen from the surface.

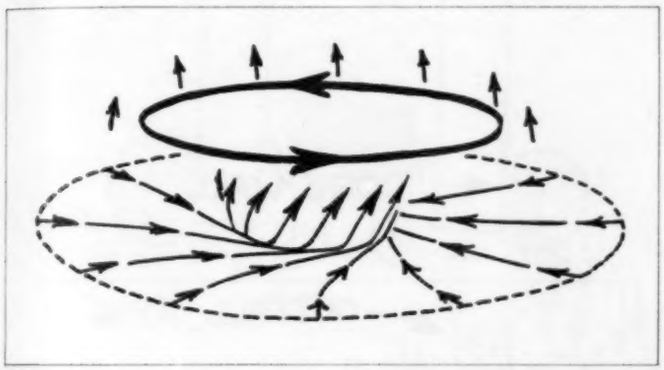


FIGURE 16.—Motions in the spray and sand bush as they would appear from an elevated station.

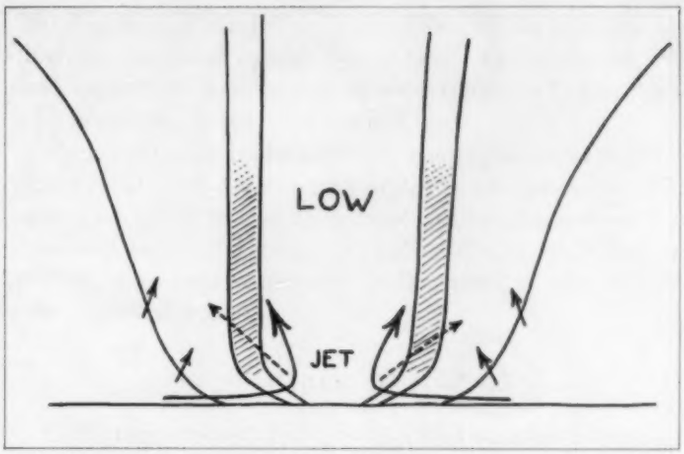


FIGURE 17.—Vertical cross section of vortex near the surface. Solid lines are schematic isobars. Shading is the "core wall" where gradient winds preclude convergence. The long arrows show the strong surface convergence under the wall, which changes to the jet upthrust inside the wall. Dashed lines show the jet cone of solid objects and particles lifted by the jet and evicted through the wall by centrifugal force. Short arrows show relatively weak updraft outside the wall.

minutes backed to ENE force 7; then the squall ended abruptly. About this time brief damaging ENE winds were felt ¾ mile SE by S of my position. This might mean another tornado. One person saw three spray bushes over the ocean, two more than I saw, so there were probably five or six disturbances in all.

At 1700 EST the cumulonimbus was rapidly dissolving, with S wind force 2, variable to W force 2, then steady SSW force 3, increasing to force 4 at 1900. Temperature and dewpoint readings taken right after the storm were lost, but the pressure then (1700) was 30.07, or plus 0.07 over that of two hours before. At 1830 temperature was 83° F., which shows little if any cooling effect after the storm, and dewpoint was 79°, slightly above the pre-storm 78°. By this time thick altocumulus covered all but the far ESE sky, and convective clouds had disappeared.

Figure 14, as mentioned earlier, shows some representative effects of the storm's action. Winds were stronger and extended farther on the right of the track, but dropped objects were found farther to the left of the track,

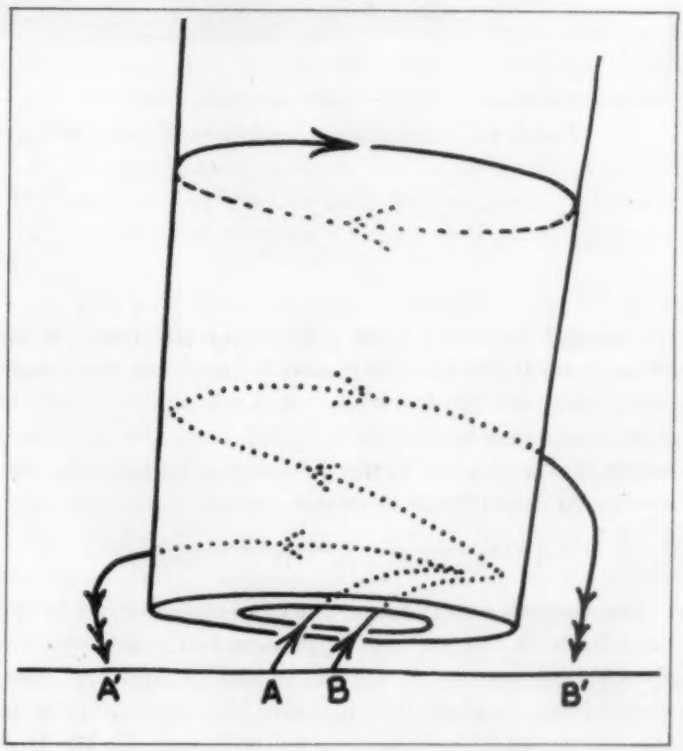


FIGURE 18.—Typical paths of objects hurled aloft by the jet; A—A' evicted through the core wall and eased to earth in the outer updraft, B—B' held captive for longer periods above center of jet.

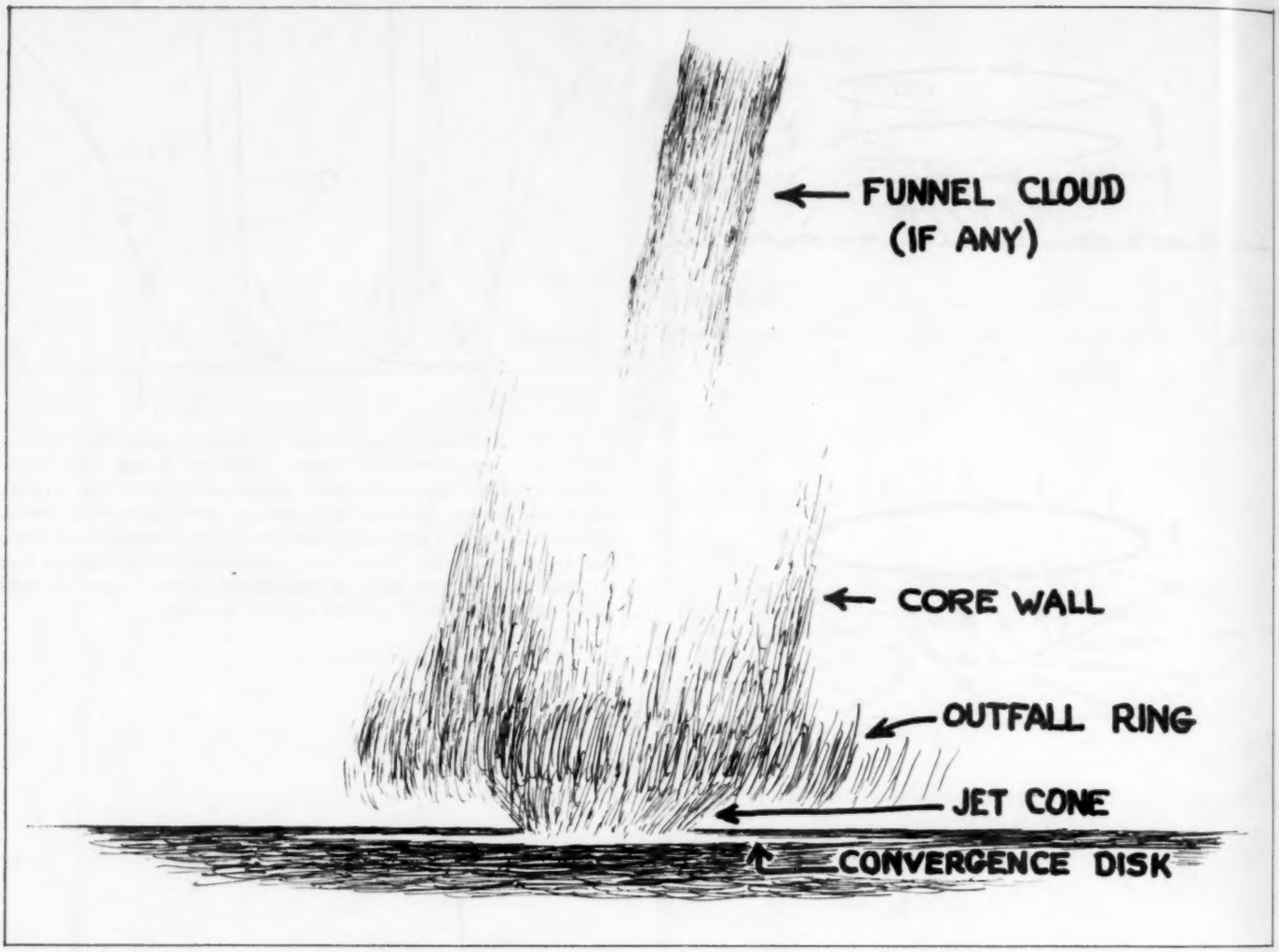


FIGURE 19.—Typical appearance of spray bush, with parts named in accordance with the author's jet hypothesis.

the overall width of their path being 600 feet. Wind effects comparable to those caused by hurricane force were seen over a path 320 feet wide. A few low pressure effects were seen, such as the door pulled from the observer's house, and a window frame and sashes pulled from the house next door toward the storm center.

4. MOTIONS IN THE TORNADO

The motions of sand and other objects observed in the sand bush as the tornado approached the observer are shown schematically in figures 15 and 16, and are interpreted in terms of air flow in figure 17. Some hints as to the movement of sand are shown in figures 11, 12a, and 12b, which are photographs of the sand bush. Horizontal circular movement 15 or more feet above the ground in what I shall call the core wall was the fastest of all observed movements in this tornado. One may hypothesize that the wind in the core wall rotates at gradient velocity and precludes convergence, except in the layer next to the ground. Here frictional drag destroys the gradient equilibrium and lets surface air rush under the wall in a converging disk and then turn suddenly upward into the low pressure at the core center. This results in a high-speed jet which apparently loses force quickly with height.

This jet is believed responsible for the prodigious feats of lifting and shattering observed in this storm, a tornado

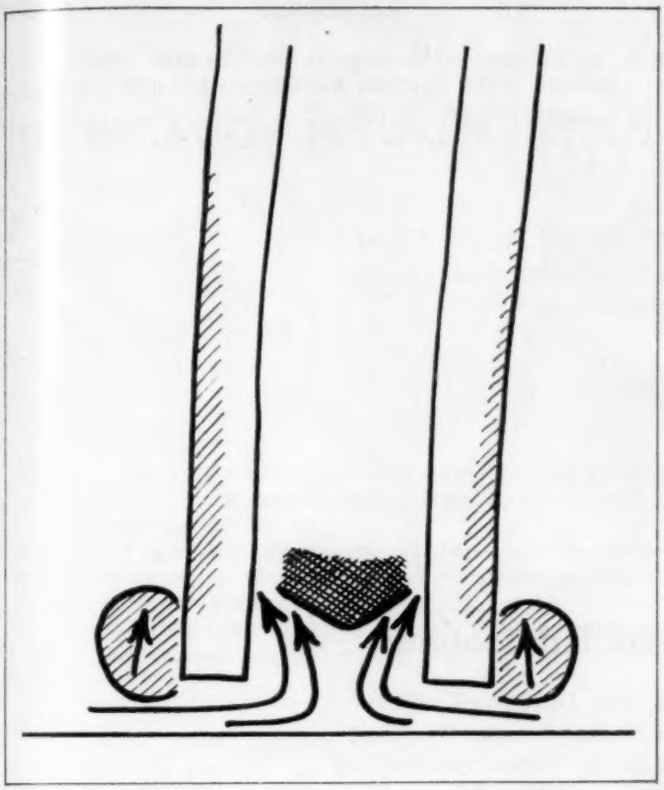


FIGURE 20.—This figure shows schematically the regions where lifted objects and particles were suspended in this tornado. Sand and spray tended to occupy the outer zones such as core wall and outfall ring (slanted shading), while heavier objects were seen suspended in the heart of the jet (the cross-hatched region). Departure of objects from these regions seemed to be slow and gradual, or even imperceptible.

of moderate intensity. Some objects lifted by the jet tended to fan out through the core walls by centrifugal force within a cone-shaped region, and when they reached the relatively mild updraft outside the wall, they formed a suspended ring and settled slowly back to earth. This sequence of movements is illustrated in figure 18. This process could also be seen in the spray bushes of the offshore waterspouts—the flat spindrift convergence disk, the spouting jet cone in the middle, the vicious whirl of the core wall above, and the gentler motion of the outfall ring. This is shown in figure 19. Figure 17 presents in vertical cross-section the schematic isobaric pattern believed to be associated with the jet phenomenon.

The core wall, about 400 feet in diameter while near the observer, made one turn in about 5 seconds, which is rotation at 175 m.p.h. Velocity in the convergence disk near the jet probably exceeded 100 m.p.h. On Engagement Hill, low-flying sand showed that the disk extended 500 feet east of the center. The effect of the convergence disk, as indicated by the flying sand, was photographed and appears on the right-hand side of figures 12 a and b as a lighter area beneath the outfall ring.

Figure 20 shows schematically the regions where lifted objects and particles were suspended in this tornado. The sand and spray tended to occupy the outer zones such as core wall and outfall ring (slanted shading), while heavier objects were seen suspended in the heart of the jet (the cross hatched region).

5. SUMMARY OF FINDINGS

1. Waterspouts of the type described here are tornadoes.

2. The formation of these tornadoes was anticipated on the basis of oppositely moving, warm, humid air streams, provided both streams could partake of convectional uplift in the same cumulonimbus cell. Warmth, humidity, and lapse rate favorable to vigorous convection in both streams seemed essential to tornadogenesis in this case. Likeness in these characteristics was more important that contrast.

3. Cyclonic circulations across the cloud base with vertical axis preceded, and gave strength to, tornado circulation of much smaller diameter which emerged from the cloud on a slant. Opposite edges of the cloud base, moving in opposite directions in tornado cyclones, probably give rise to frequent eyewitness reports of "two clouds coming together."

4. There was no evidence of downrush or outflow at the surface, and only isolated evidence of downrush aloft. The evidence favored the opposite—convergence and uprush at the surface.

5. Strong convergence at the surface was converted suddenly to a vertical jet near the core center. The converging air acquired rotary motion as it passed under the core wall.

6. The jet lifting effect was tremendous near the surface, but relatively weak a short distance aloft. Other tornadoes have, of course, lifted objects high into the clouds.

7. The lower portions of tornadoes 2 and 3 did not, while slanting, trail behind the funnel clouds aloft, but preceded them without help from the surface wind.

8. The funnel clouds lowered, rose, broke, and joined capriciously and with no apparent relation to the surface violence of the storm which in this instance raged steadily on until the final decline.

9. The funnel cloud need not come anywhere near the ground for the tornado to develop destructive power at the surface.

ACKNOWLEDGMENT

The author acknowledges the kindness of the late Dr. C. F. Brooks and of his son, Dr. E. M. Brooks, who reviewed the manuscript of this paper and offered many substantial suggestions for its improvement; and the kindness of Mr. Walter Hoecker, Jr., for his assistance in editing the manuscript and the illustrations.

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New Weather Bureau Publications

Technical Paper No. 29, "Rainfall Intensity-Frequency Regime: Part 4—Northeastern United States," Washington, D.C., 1959, 35 pp; for sale by Superintendent of Documents, U.S. Government Printing Office, Washington 25, D.C., price $1.25.

Paper analyzes rainfall intensity-duration-frequency regime, with other storm characteristics, for duration of 20 minutes to 24 hours, areas from point to 400 square miles, and frequencies for return periods from 1 to 100 years, for the region east of longitude 80° W. and north of latitude 41° N.

THE RELATION BETWEEN HEIGHT PATTERNS AT 500 MB. AND 100 MB.

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[Manuscript received June 5, 1959; revised July 24, 1959]

ABSTRACT

The flow at 500 mb. is compared with that at 100 mb. With the exception of the polar vortex, the patterns at the two levels tend to show a distinct similarity, with the major exception that small-scale features at 500 mb. are not present at 100 mb.

In middle latitudes, the wind speeds at the two levels are of the same order of magnitude, with a systematic excess of the 100-mb. speeds over the 500-mb. speeds at latitudes 30° and farther south.

In the region of the polar vortex, heights at the two levels are still well correlated. Variations in height over long periods tend to be much larger at 100 mb. than at 500 mb. Similarly, the winds at 100 mb. are much stronger than those at 500 mb.

1. INTRODUCTION

In middle latitudes, the intensity of the winds generally increases through the troposphere and decreases above. It might be expected, therefore, that there would correspond to a given level in the troposphere, a level in the stratosphere at which the wind speeds are the same. This hypothesis is based, in part, on the well-known fact that, in middle latitudes, the change from a warm to a cold upper troposphere at a given station is compensated by a change from a cold to a warm lower stratosphere and vice versa.

Since the 500-mb. level has become the standard level for research and forecasting in the middle troposphere, a "matching" level to 500 mb. in the stratosphere might prove convenient. Fortunately, in middle latitudes, such a "matching" level is often not far from 100 mb., which itself is a mandatory surface.

Of course, the flow at 500 mb. and 100 mb. may be different for several reasons. In the first place, the troughs and ridges with shorter wavelengths at 500 mb. are damped upward and do not appear at 100 mb. This is brought out by the difference in the spectra of height or winds at two levels, which show in every case that the 500-mb. flow contains more high frequency variations compared to low frequency variations than does the 100-mb. flow. Because of this difference between the levels, similarity between the flows at the two levels is enhanced if the flow at 500 mb. is smoothed in some manner. Some experimentation has indicated that the 20° diamond space-mean at 500 mb. (Berry, Haggard, and Wolff [1]) is most satisfactory for this purpose.

Similarly, if the 500-mb. heights or winds are averaged over a time interval equal to or greater than the period of the systems to be averaged out, the two levels can be compared conveniently. In practice, 4-day averages have been employed at both levels in some of the studies presented in this paper; such averages tend to eliminate small-scale flow at 500 mb. and errors at 100 mb.; further, such averages can be estimated at 100 mb. even if a portion of the observations is missing.

Exact equality between smoothed 500-mb. charts and 100-mb. charts can be expected only if tropospheric warming and stratospheric cooling between these two levels exactly compensate. There are several reasons why such compensation may not occur: in the Tropics, the tropopause may be near 100 mb., so that the temperature changes between 500 mb. and 100 mb. are almost entirely tropospheric. In those cases, the wind, on the average, increases from the 500-mb. level up to near 100 mb., so that the winds at 100 mb. exceed the winds at 500 mb.

In contrast, at high latitudes, the tropopause tends to be quite low, so that, if there is stratospheric and tropospheric compensation, the winds at 100 mb. are slower than those at 500 mb. However, it will be seen later that compensation between stratospheric and tropospheric temperatures does not always occur. If the region in question is influenced by the "polar regime" (to be defined later), temperature changes tend to be of the same sign at all levels, and wind changes, as well as the winds themselves, are much greater at 100 mb. than at 500 mb. Hence, the comparison between smoothed 500-mb. flow and 100-mb. flow will indicate what regime is dominant over a given region.

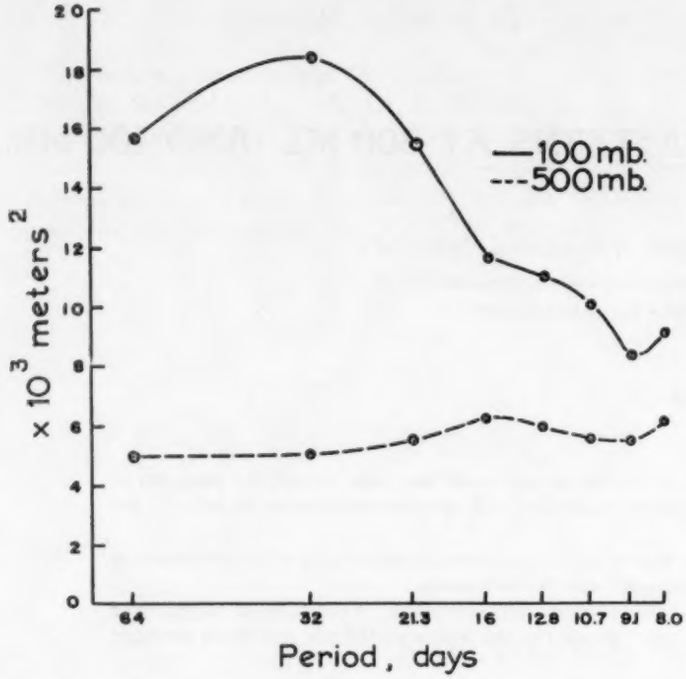

FIGURE 1.—The 100-mb. and 500-mb. height spectra within the polar regime. Data are averages from selected Canadian stations: Frobisher, Arctic Bay, Coral Harbour and Baker Lake for winter 1953–54; and Alert, Port Harrison, Resolute, and Baker Lake for winter 1956–57.

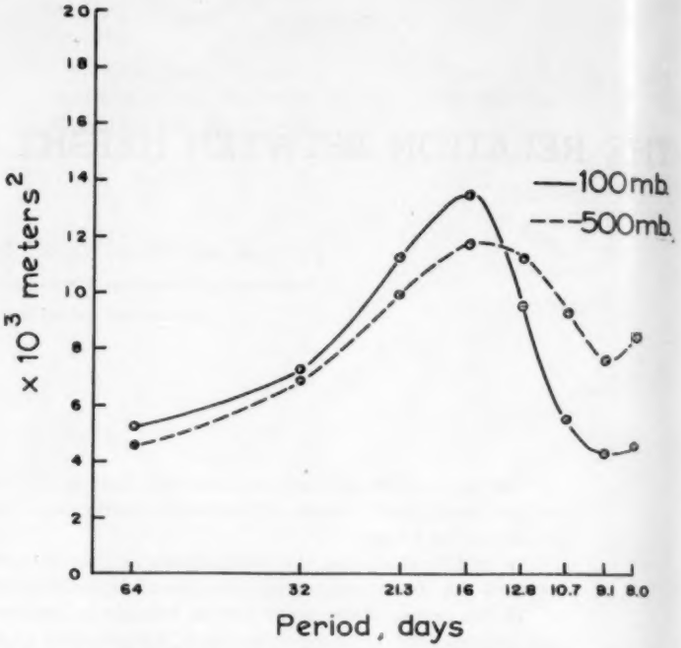

FIGURE 2.—The average 100-mb. and 500-mb. height spectra outside the polar regime, for winters 1955–56 and 1956–57 for Chaumont, France and Nome and Annette Island, Alaska.

2. CLASSIFICATION OF REGIMES AT 100 MB.

Throughout much of the lower stratosphere, contour heights decrease toward the pole and temperatures increase toward the pole. This behavior is not found in the upper stratosphere, above the 50-mb. level. For this reason, the regime in which the temperature increases poleward will be referred to as LSR (lower-stratosphere regime). The LSR extends from about 30° latitude to the poles in summer, and to about 55°, on the average, in winter. Since contour height gradients and temperature gradients oppose each other, the wind speed decreases upward in the LSR. Further, the LSR is characterized by warm troughs and cold ridges, which also decrease upward in intensity. In other words, patterns in the LSR increase in intensity downward, and are, presumably, most strongly developed near the tropopause; therefore, they are closely related to tropospheric patterns.

Within the 55° latitude circle in winter on the average, temperatures and heights both decrease poleward; westerly winds increase upward in intensity. This regime, called the PR (polar regime), is characterized by warm Highs and cold Lows. Its main feature is usually a cold Low, increasing upward in intensity, which will be called the "polar vortex." From studies of the radiation budget of the stratosphere by Ohring [8], the existence of this vortex can be said to be a result of strong radiational cooling of the stratosphere in the region of the polar night. The reader is referred to Kochanski [7], Godson and Lee [4], and Godson et al. [3] for further details.

Between the PR and the LSR is located the PRB (for polar-regime boundary). This is a band of relatively high temperatures. It is, however, not necessarily the warmest region at 100 mb. Frequently, warm regions associated with cold tropospheric Lows occur in these latitudes. The heights are relatively low so that we deal with warm Lows characteristic of the LSR. Rather, the PRB must be defined as a rather narrow circumpolar region which separates systems intensifying upward from systems decreasing upward. Further, it separates regions in which local temperature and height changes have the same sign (PR) from regions where they have opposite sign (LSR). Finally, winds increase upward in the PR and decrease upward in the LSR. Theoretically, any one of these criteria can be used to locate the PRB. In practice, the easiest method seems to be a careful isotherm analysis indicating the transitions from cold to warm ridges and warm to cold troughs.

Neither the LSR, PR, nor PRB remain symmetrical with respect to the pole. The center of the polar vortex at 100 mb. has been observed to move 20° away from the pole and the PRB to extend down to latitude 40°. The polar vortex center tends to drift very slowly, may take quite different positions in different winters, and appears to

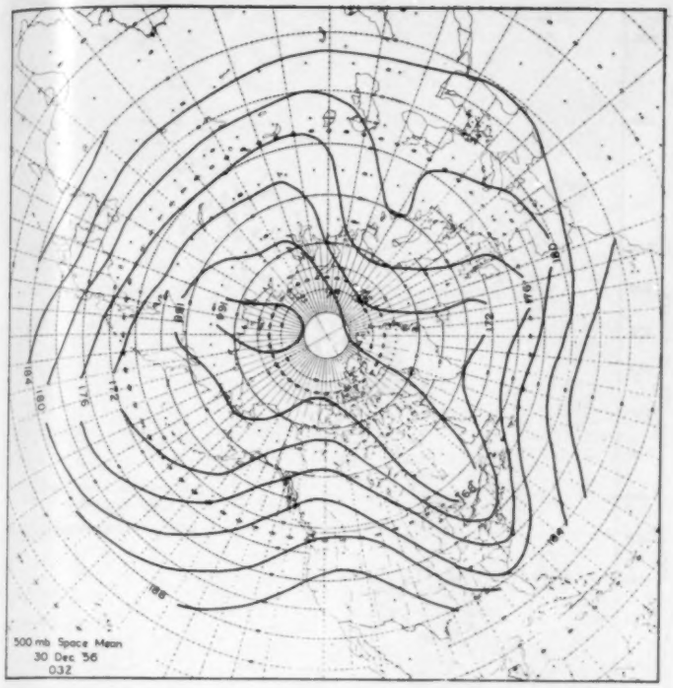

FIGURE 3.—20° diamond space-mean for 500 mb., 0300 GMT, December 30, 1956. Contours labeled in hundreds of feet.

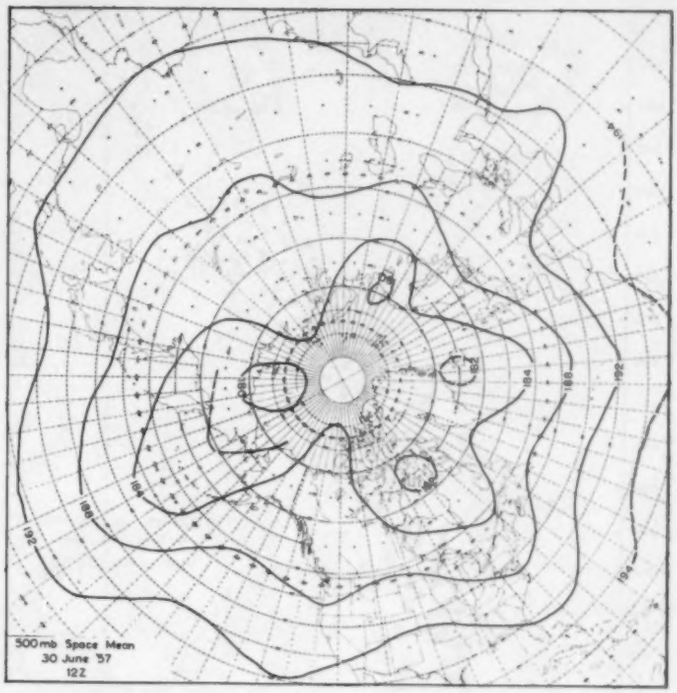

FIGURE 5.—20° diamond space-mean for 500 mb., 1200 GMT, June 30, 1957. Contours labeled in hundreds of feet.

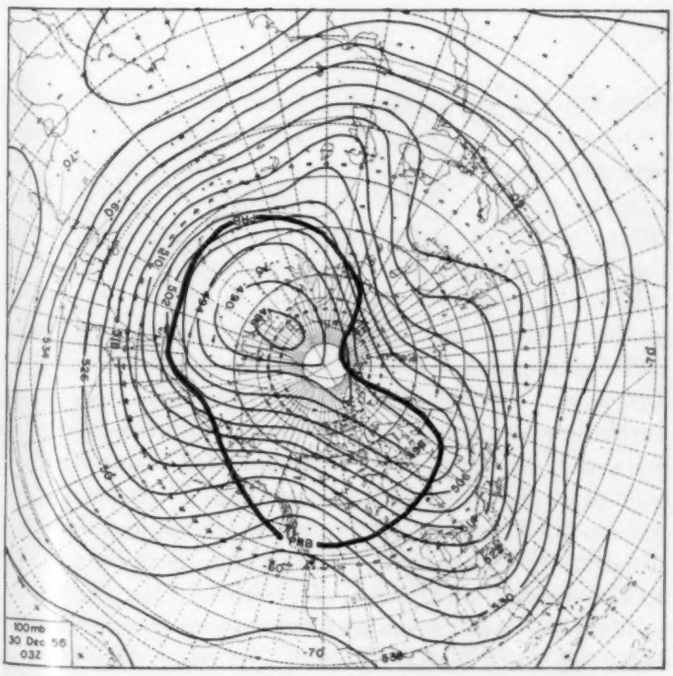

FIGURE 4.—100-mb. chart, 0300 GMT, December 30, 1956. Contours labeled in hundreds of feet. Polar regime boundary is indicated.

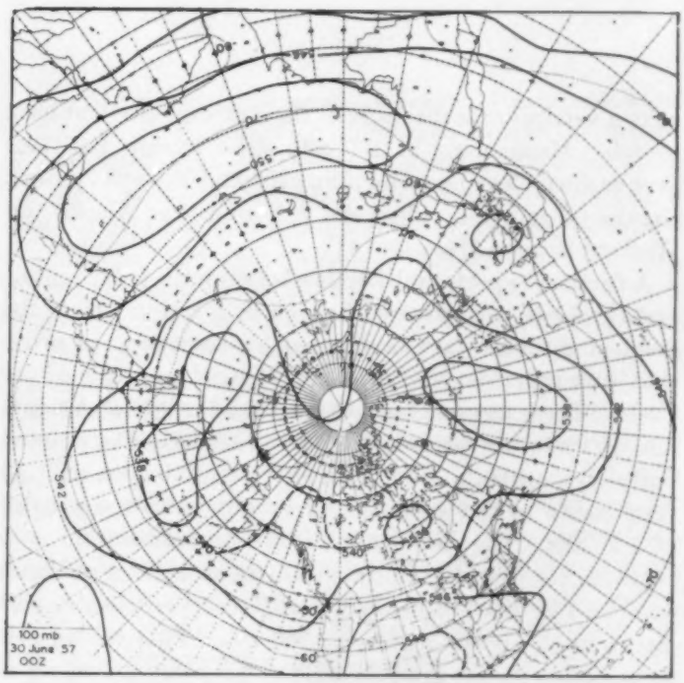

FIGURE 6.—100-mb. chart, 0000 GMT, June 30, 1957. Contours labeled in hundreds of feet.

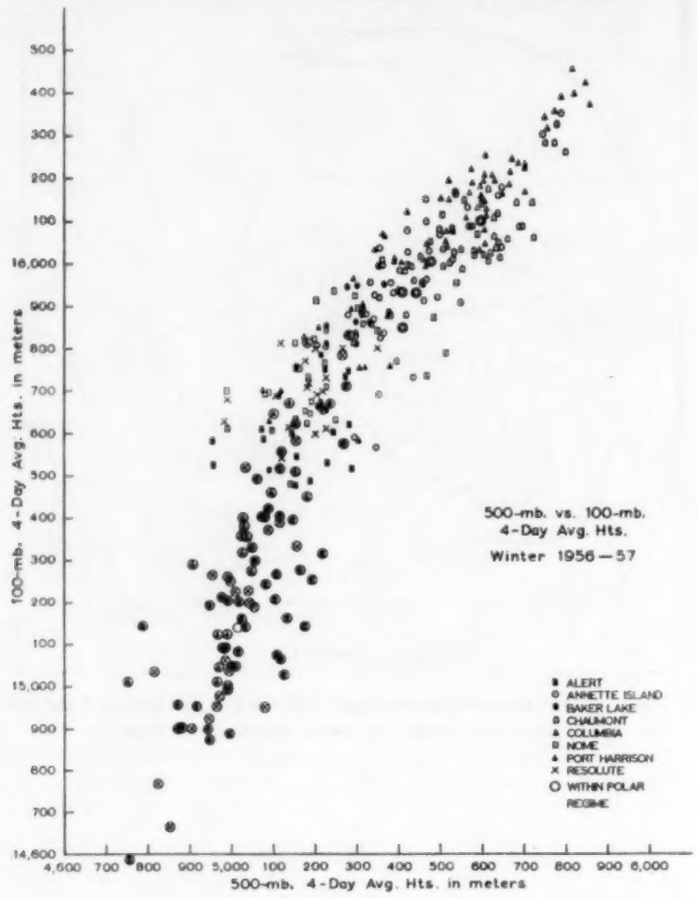

FIGURE 7.—Four-day average 100-mb. heights versus four-day average 500-mb. heights for winter 1956–57. The symbols representing periods in the polar regime are circled.

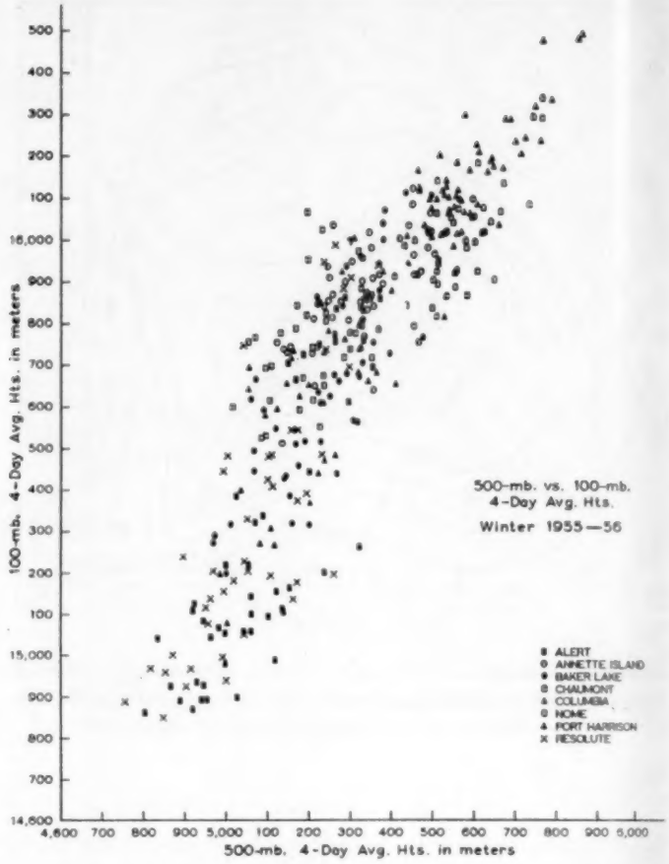

FIGURE 8.—Four-day average 100-mb. heights versus four-day average 500-mb. heights for winter 1955–56.

have a vertical axis throughout the stratosphere to at least 10 mb. Mean maps at 100 mb. show trough positions over eastern Canada, eastern Siberia, and Novaya Zemlya. On the other hand, ridges are indicated over Alaska and Western Europe (Heastie [5]).

A further important difference between the polar regime and the LSR is illustrated by figures 1 and 2. Here the spectral estimates are multiplied by frequency in order that area represent variance on a logarithmic scale of periods. These figures contrast spectra of height at 500 mb. and 100 mb. on the two sides of the PRB. As pointed out above, on both sides the relative contribution of high frequencies to the variances of height is always greater at 500 mb. The important difference between the two regimes, however, lies in the enormous amount of variance produced by low frequency fluctuations (periods above 20 days) in the polar regime at 100 mb. This variance is associated with the gradual drift of the polar vortex noted above, or with slow deepening or filling of the vortex. The strong baroclinity associated with the stratospheric cooling in the polar night is a logical source of this energy.

Outside of the polar vortex, however, the difference between the spectra at 500 mb. and 100 mb. can be understood by assuming that the energy is created in the troposphere and damped upward above the tropopause, with the damping being most effective at high frequencies.

The polar regime usually begins to form in September as the strong radiative cooling of the polar night begins and continues to increase in intensity until the end of December. Its disappearance may be quite sudden, occurring some time between the beginning of January and the end of March in the winters thus far examined (1950 to 1958). The date of the beginning of this breakup, as well as the intensity of the breakup, is quite different in different winters. Further, each winter seems to be unique in the strength and preferred position of the vortex.

Many of these phenomena have also been described by Scherhag [10], Godson et al. [3], Godson and Lee [4], Teweles [11], Teweles and Finger [12], Craig and Hering [2], and Warnecke [13].

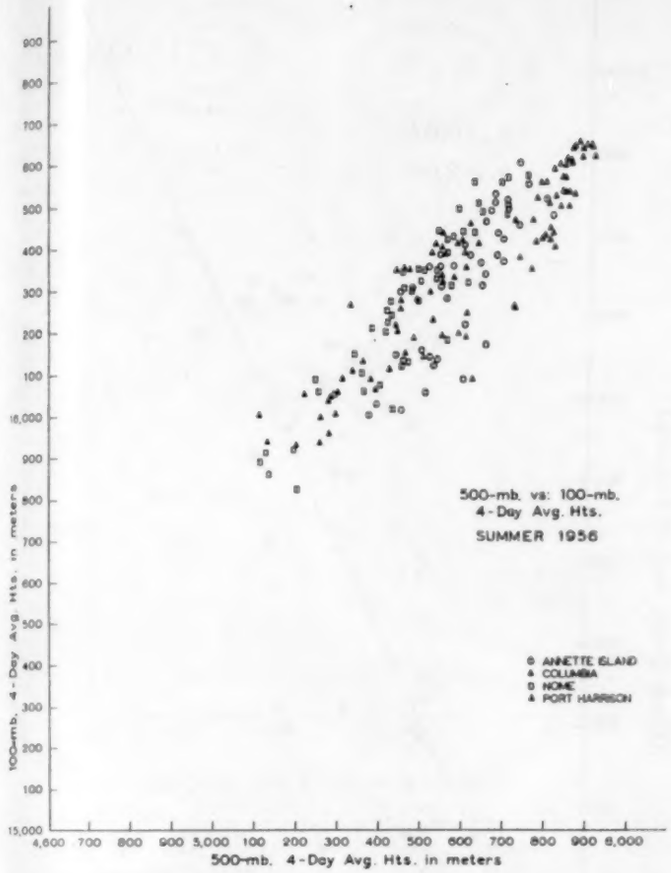

FIGURE 9.—Four-day average 100-mb. heights versus four-day average 500-mb. heights for summer 1956.

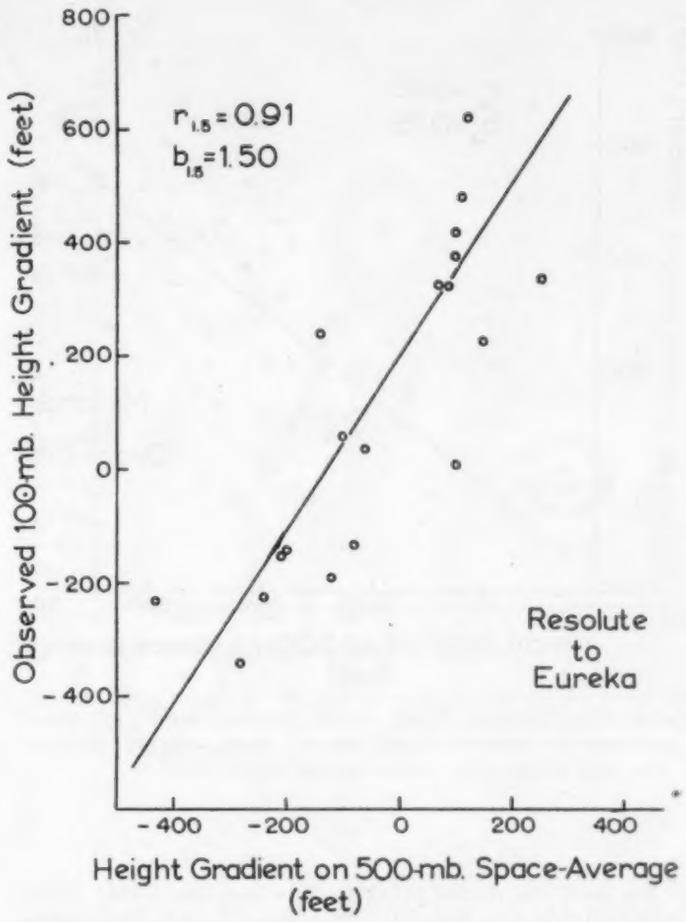

FIGURE 10.—Observed 100-mb. height gradient versus the height gradient on the space-averaged 500-mb. chart between Resolute and Eureka, Canada during winter 1956–57.

3. COMPARISON BETWEEN SPACE-MEAN 500-MB. FLOW AND 100-MB. FLOW

Figures 3 and 4 compare the 500-mb. height distribution, smoothed by the AROWA 20° diamond space mean (see Berry, Haggard, and Wolff [1]), for 0300 GMT, December 30, 1956, with the 100-mb. height distribution. In many ways this situation is typical for a winter with a well developed vortex. The 100-mb. flow shows a vortex with wave numbers two and three predominating; also the PRB is located in figure 4.

Equatorward of the PRB the 100-mb. flow shows a remarkable similarity to the space-mean 500-mb. flow with, of course, the horizontal temperature gradient reversed. Between latitude 35° and the PRB, the wind speeds are about equal at the two levels. South of 35° there is a tendency for the 100-mb. speeds to exceed those at 500 mb. at least in the areas of adequate data. In all cases, the trough-ridge positions at the two levels match as well as can be expected. This means that wind directions on the two charts are well correlated.

A study involving three independent 10-day periods in each of three winters was undertaken to test the degree of correspondence of the space-averaged 500-mb. and 100-mb. charts. Geostrophic winds were evaluated at a series of grid points covering the United States and southern Canada, and root-mean-square speed and direction differences formed. At all grid points root-mean-square direction differences averaged 19 degrees, varying over the periods involved from 11 to 25 degrees. The speed differences were strongly a function of latitude; at 30° the speeds averaged consistently 15 knots greater at 100 mb. than at 500 mb. At 40°, the differences averaged from 0 to 27 knots, and at 50°, from 4 to 29 knots. At 40° and 50° the speed differences are strongly affected by the degree of strength of the stratospheric vortex circulation. At 30° the tropopause is near 100 mb., no temperature compensation exists, and the 100-mb. speeds are correspondingly greater. At 40° the matching level is near 100 mb., if the polar vortex does not affect this latitude.

Perhaps rather unexpectedly, wind directions seem to be matched well inside the PRB. Of course, the spacing

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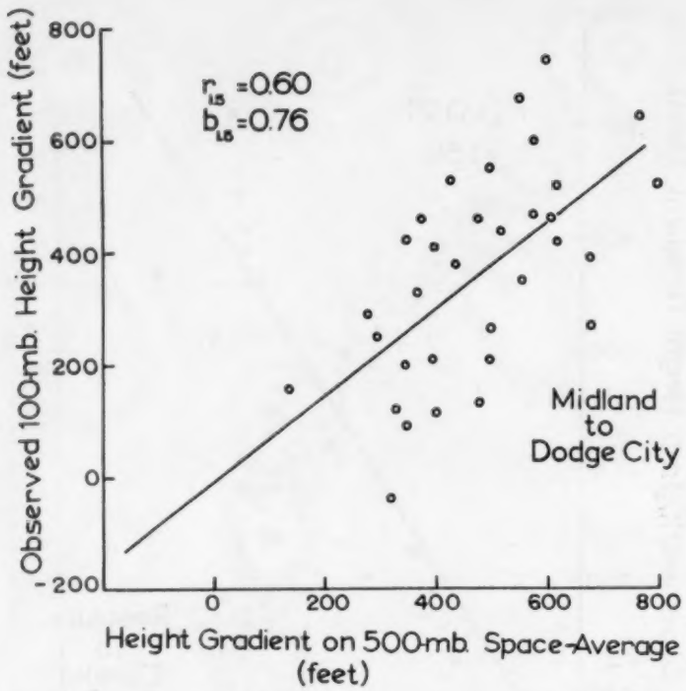

FIGURE 11.—Observed 100-mb. height gradient versus the height gradient on space-averaged 500-mb. chart between Midland, Tex. and Dodge City, Kans. during winter 1956–57.

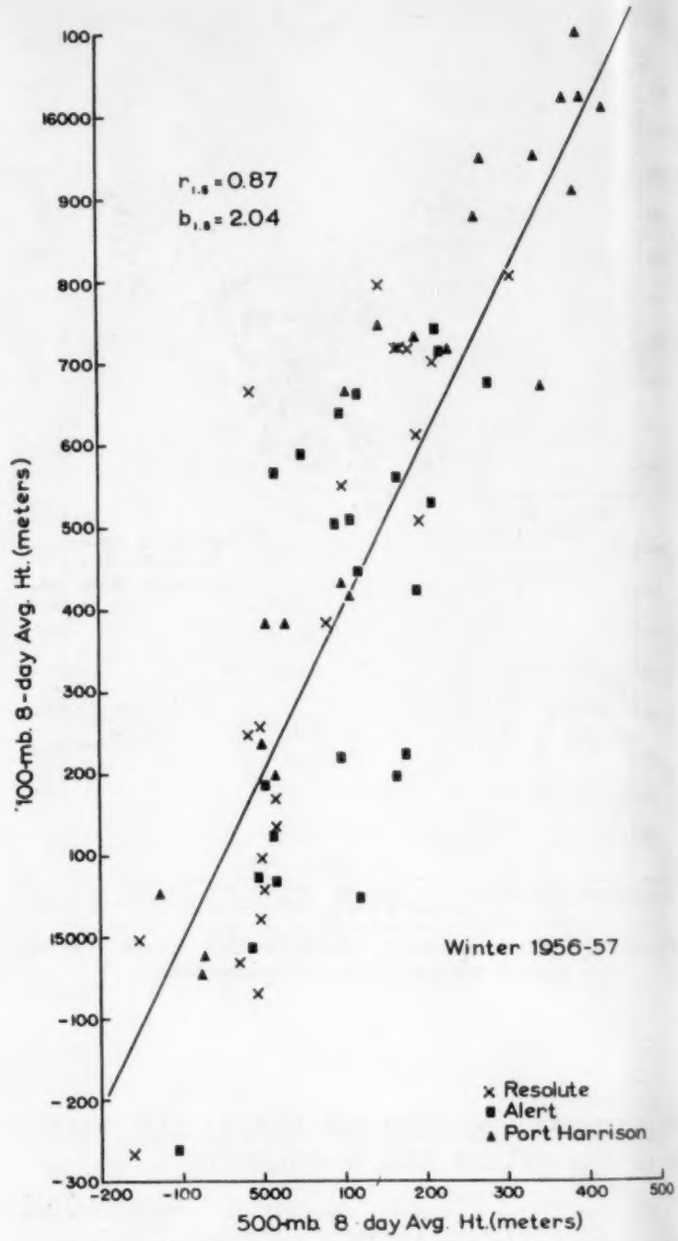

FIGURE 12.—Eight-day average 100-mb. heights versus eight-day average 500-mb. heights within the polar regime during winter 1956–57.

of the contours in the polar regime is considerably closer at 100 mb. than at 500 mb. This more intense circulation in the stratospheric polar circulation in winter will be analyzed in greater detail later.

Figures 5 and 6 compare a 500-mb. space-mean chart in summer, June 30, 1957, with the corresponding 100-mb. chart. The tropospheric systems (in middle and high latitudes) are quite recognizable at 100 mb., although the speeds at the higher level are generally weaker than the space-mean gradients. Examination of vertical wind speed profiles indicates the "level of match" to be about 125 mb. at latitude 40° N. lowering to 150 mb. at latitude 60° N. Again, emphasis must be placed on recognition that the matching level idea is restricted to regions in which temperature compensation occurs between the upper troposphere and lower stratosphere. Thus the subtropical Highs and regions wherein the tropical tropopause is present cannot be expected to show such characteristics. This can easily be seen by examining wind profiles in the southern United States for the June 30, 1957 maps shown.

4. DIFFERENCES IN THE STRATOSPHERE-TROPOSPHERE RELATIONS ON THE TWO SIDES OF THE PRB

For eleven stations (1 in Missouri, 1 in France, 7 in Canada, and 2 in Alaska), 4-day average heights at 500 mb. were compared with simultaneous 4-day average heights at 100 mb. These quantities will be simply referred to as "heights." The observations covered four winters (defined as the period from October through March), from the 1953–54 season to the 1956–57 season. The remaining periods of the years 1955, 1956, and 1957 were defined as "summer" and analysed separately.

Figure 7 shows the relationship betwen 500-mb. and 100-mb. height for stations which were on different sides of the polar regime boundary (PRB) at different times. Points representing observations at times when the station was inside the PRB are circled. A station was de-

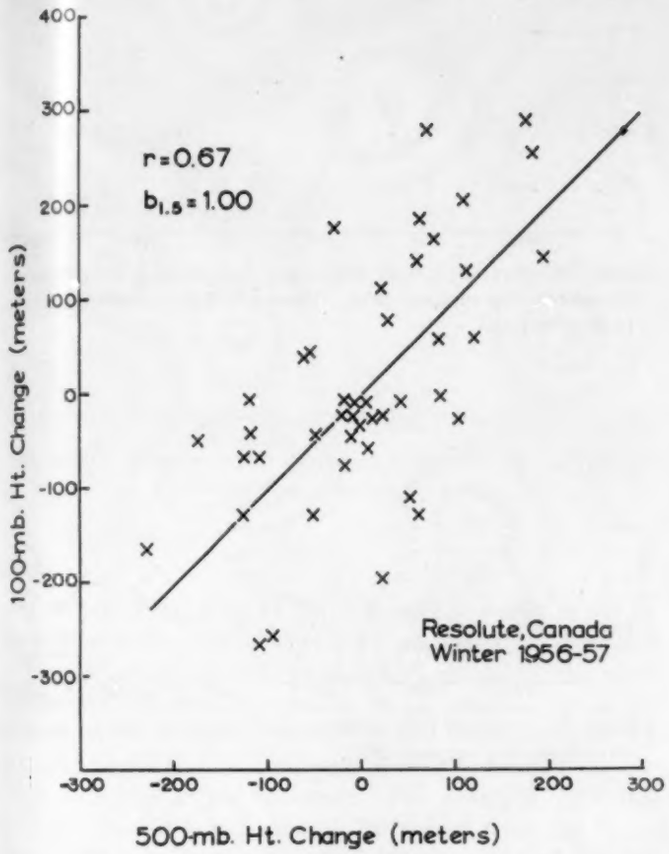

FIGURE 13.—Change in four-day average 100-mb. height versus the change in 4-day average 500-mb. height at Resolute, Canada during winter 1956–57.

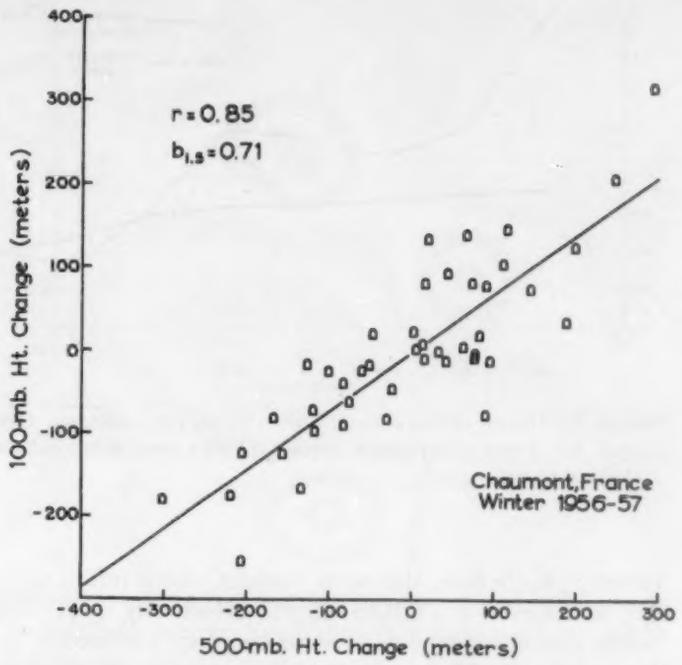

FIGURE 14.—Change in four-day average 100-mb. height versus the change in four-day average 500-mb. height at Chaumont, France during winter 1956–57.

fined to be within the PRB when its height tendency and temperature tendency had the same sign. The figure suggests generally that there is a good relation between the heights at the two levels.

Figure 7 serves to illustrate a basic difference between the stratosphere-troposphere height relationships on the two sides of the PRB: outside the PRB, the slope of a line of regression for 100-mb. height as function of 500-mb. height does not differ much from unity, indicating stratosphere-troposphere temperature compensation; inside the PRB, in the polar regime, the slope is considerably greater than unity, showing that compensation is absent or incomplete. Figure 8 illustrates the same features for another winter season. Figure 9, on the other hand, was obtained from summer observations. In summer, no polar regime exists and all observations are consistent with the hypothesis that the relation between heights at the two levels has a slope of about unity in the LSR.

That a similar difference in the winds exists in the PR and LSR was already made plausible in the last section; this point is further illustrated by figures 10 and 11. These figures contrast the relation between geostrophic wind components at 100 mb. and 500 mb. in a region dominated by the polar regime and in a low-latitude region. Again, the wind fluctuations in the polar regime are greater at 100 mb. than at 500 mb., but of the same general magnitude outside of the PRB.

It was shown in section 2 that the spectra of the heights inside the polar regime in the stratosphere are characterized by an enormous amount of low-frequency variance corresponding to periods greater than 20 days. The question may be raised whether the large ratio of 100-mb. variations to 500-mb. variations is associated with these large and gradual variations. If that were true, one would expect that the slow fluctuations of height at 100 mb. would be of a much greater magnitude than those at 500 mb., whereas the high-frequency behavior at the two levels would be similar. That this is indeed the case is demonstrated by figures 12 and 13.

Figure 12 shows the relation between 8-day mean heights at 500 mb. and 100 mb. The effect of taking 8-day means is to filter out practically all energy of height fluctuations with periods less than 8 days, and more than one-half the energy of oscillations of the order of 16 days. In other words, the bulk of the variation indicated by figure 12 is due to fluctuations with periods exceeding 16 days. Apparently, the correlation between 500 mb. and 100 mb. is quite large, but the magnitudes of the changes at 100 mb. are more than twice those at 500 mb.

Figure 13 shows the relation between height changes at 500 and 100 mb., rather than heights, inside the polar

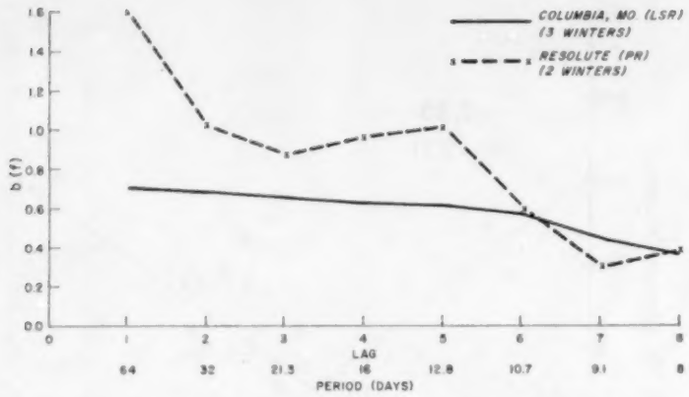

FIGURE 15.—Ratio of cospectrum (500–100 mb.) to spectrum (500 mb.) for lower-stratosphere regime (LSR) and polar regime (PR) stations.

vortex. As before, the term "height" here refers to 4-day averages. It can be shown (Holloway [6]) that height changes (as compared to heights) are mostly influenced by the high-frequency components of the height spectra. Of course, these changes are also influenced relatively strongly by errors of observation, a fact which explains at least part of the large scatter evident in figure 13. Nevertheless, the figure makes plausible the conclusion that, in the polar vortex region, height fluctuations with periods of 8 days or less at 100 mb. and 500 mb. are of the same general magnitude.

Outside the polar regime, even the slow height fluctuations are of the same order of magnitude at 500 mb. and 100 mb. This fact is demonstrated by figure 14. It is reasonable to assume that these variations are greatest at about the level of the tropopause, and thus of about the same magnitude at 500 mb. and 100 mb. The compensation between stratospheric and tropospheric temperature changes here is good.

The fact that the relationship between 500 mb. and 100 mb. in the PR depends on scale was further confirmed by the cross-spectrum analysis between the heights at the two levels. We can define a quantity $b(f)$ by:

$$b(f)=\frac{C(f)}{S(f)};$$

here $C(f)$ is the cospectrum between 500-mb. and 100-mb. heights, and $S(f)$ is the spectrum of 500-mb. heights. The quantity $b(f)$ can be interpreted as the regression coefficient at frequency f, which would be needed to estimate a given harmonic at 100 mb. from the same harmonic at 500 mb. As figure 15 shows, $b(f)$ decreases with increasing frequency, as would be expected from the arguments given above.

5. THE VARIATIONS OF WINDS AT 500 MB. AND 100 MB.

The influence of the position of the PRB on the relation between 500-mb. and 100-mb. flow is further illustrated by figures 16–20. Here the 8-day mean height gradient between Columbia, Mo., and Resolute Bay is traced through two summers and three winters. This quantity of course, gives an average geostrophic wind component at right angles to the line connecting the two stations.

There is not much difference between the two summer seasons; the variations in height difference at 500 mb. and 100 mb. are extremely well correlated with each other. Numerically, the height gradient at 100 mb. averages about

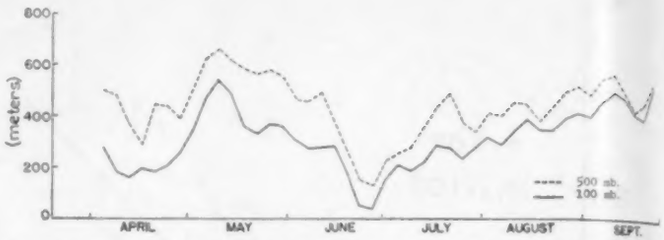

FIGURE 16.—Contour height difference, Columbia, Mo. to Resolute, Canada during summer 1956. These are 8-day running averages (4-day overlap).

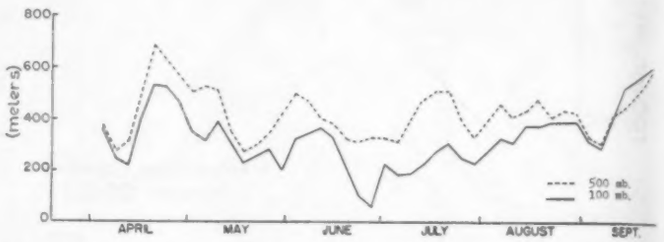

FIGURE 17.—Contour height difference, Columbia, Mo. to Resolute, Canada during summer 1957.

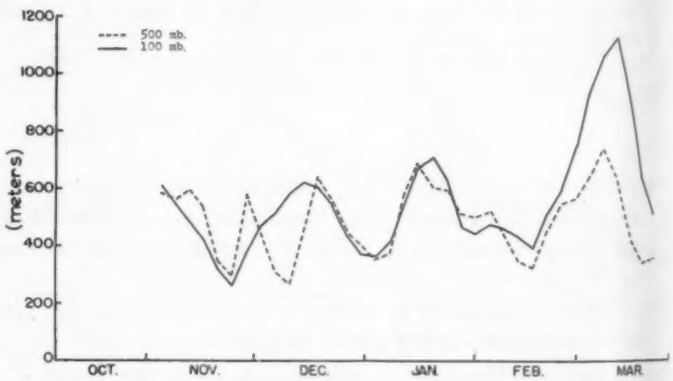

FIGURE 18.—Contour height difference, Columbia, Mo. to Resolute, Canada during winter 1952–53.

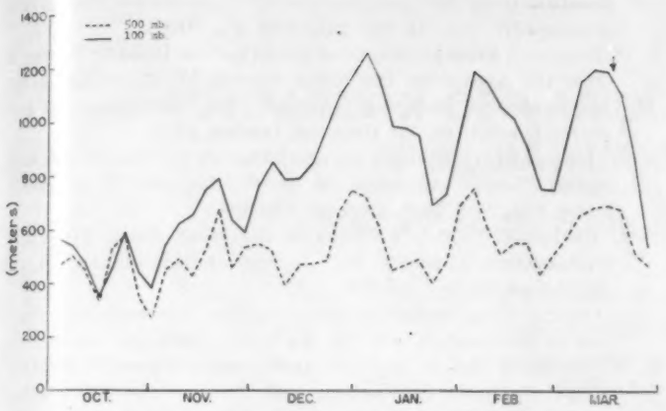

FIGURE 19.—Contour height difference, Columbia, Mo. to Resolute, Canada during winter 1955–56.

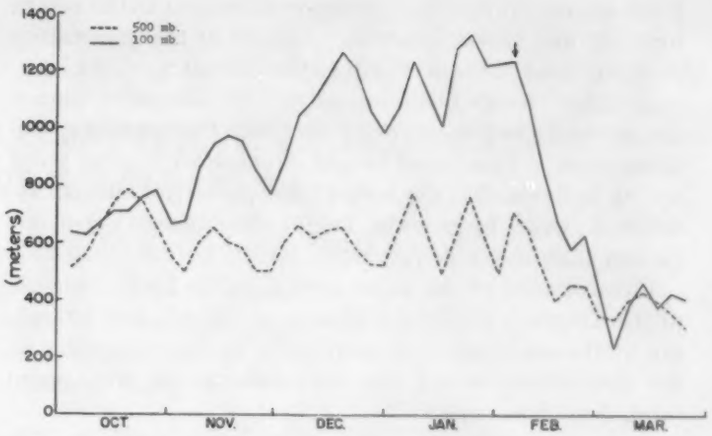

FIGURE 20.—Contour height difference, Columbia, Mo. to Resolute, Canada during winter 1956–57.

100 m. less than that at 500 mb. This result is not inconsistent with the observations discussed in section 3.

The behavior of the geostrophic winds in the three winter seasons is more varied: in all seasons, again, the changes of wind with time at the two levels are well correlated; however, the difference in wind speed at the two levels is by no means constant. For example, in winter 1955–56, the difference steadily increased from the beginning of November to the end of December (with the 100-mb. flow being the faster), then remained about constant until the second half of March when it dropped rather suddenly to zero. In 1956–57 a similar build-up of difference in wind occurred, with a sudden drop at the beginning of February. Only in the winter 1952–53 did the difference remain small, except for a short period at the beginning of March.

The differences among the winds in the various seasons are clearly related to the differing behavior of the polar vortex. For, as pointed out previously, the polar vortex results in relatively faster winds at 100 mb. than at 500 mb. Thus, for example, the polar vortex often makes its appearance in the North American section at the beginning of November, builds up to maximum intensity at the end of December, and then weakens rather suddenly at some time between January and March. In 1957, the polar vortex weakened rapidly during the time of the frequently-studied period of "explosive" warming at the end of January and beginning of February; in 1956, however, the vortex remained strong through March. For January, February, and March, 1953, we fortunately have the U.S. Weather Bureau series of daily 100-mb. and 50-mb. charts, constructed by Moreland and Cluff [9]. These maps, although dependent almost wholly on extrapolated data in the polar regions, show that in this season the vortex center was relatively far from the pole, on the Eurasian side, so that there was no large effect in the North American sector. Only in the beginning of March, when the vortex slowly drifted across the pole, did its influence become apparent over North America for a week or two before the usual weakening of early spring.

Of course, even when the influence of the polar vortex is strongest, the 100-mb. average wind speed between Columbia and Resolute Bay is not quite twice the wind speed at 500 mb., as might have been expected from the previous discussion. The reason is that only a variable amount of the northern portion of the line from Resolute Bay to Columbia is affected by the circulation around the vortex.

To summarize these results, the relation between 500-mb. and 100-mb. flow in winter north of 35° N. is dominated by the position and intensity of the vortex. When the vortex is absent, 500-mb. space-mean winds are equal to or even faster than 100-mb. winds; on the other hand, if influenced by the vortex, the 100-mb. winds are, by far, faster than 500-mb. winds.

6. SUMMARY

Examination of numerous simultaneous 100-mb. and 500-mb. charts has indicated that the basic long-wave structure of the upper troposphere is preserved in the lower stratosphere to at least 100 mb. In middle latitudes cooling below the tropopause level tends to be compensated by warming in the lower stratosphere, and vice versa. Removal of the shorter wavelength disturbances associated with strongly baroclinic systems at 500 mb. by a suitable space-mean process produces a reasonable mid- and high-latitude 100-mb. flow pattern in all seasons of the year, with some exceptions.

The major exception occurs in winter when strong radiative cooling of the stratosphere produces a cold vortex circulation which reverses the usual poleward temperature gradient in polar latitudes and prevents troposphere-stratosphere compensation. Strong westerlies are thus to be expected at 100 mb., being roughly twice the speed of the 500-mb. space-mean winds in comparable latitudes.

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Each winter exhibits uniqueness with respect to the vortex intensity and vortex location. In spite of the uncertainty thus produced in the strength of the 100-mb. flow, the long-wave ridge-trough positions, as well as the vortex center, are generally well specified by means of the 500-mb. space-mean chart. Spectra of height changes within the polar vortex indicate that the larger changes at 100 mb. are associated with large-scale height fluctuations (periods greater than about 20 days).

Equatorward of the polar vortex, in the LSR, (outside of the Tropics) the height changes at 500 mb. and 100 mb. are of the same order of magnitude, and the intensity of the circulation, which has its maximum at tropopause level, diminishes upward.

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COMPARISON OF VARIOUS CONDITIONING METHODS FOR RAWINSONDE BALLOONS

Results From the South Pole Station, Antarctica, 1957

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Manuscript received May 12, 1959; revised July 10, 1959

ABSTRACT

Various methods of conditioning the rawinsonde balloon were tried at the Amundsen-Scott IGY Station (South Pole), Antarctica, in an attempt to improve sounding heights. During the dark, cold months, conditioning of the balloon in warm diesel fuel proved significantly better than the conventional conditioning methods. However, with the return of sunlight and the rapid warming of the stratosphere, diesel fuel conditioning was found to be inferior to the boiling water or moist heat-box method of balloon conditioning.

1. INTRODUCTION

With the advent of the International Geophysical Year, various international organizations (World Meteorological Organization, Special Committee for the IGY) voiced the request that all meteorological services attempt to obtain radiosonde flights to 50 mb. on a regular basis. At most of the United States installations this did not present a problem as it was already being accomplished. However, for the United States IGY stations being established in Antarctica it did present an operational problem. All of the evidence available from radiosonde observations made in the Antarctic indicated that the stratosphere over the continent becomes extremely cold during the winter season (approaching −90° C. at 18–20 km.). The deleterious effect of these cold temperatures on the neoprene sounding balloon as reflected in much lower flight altitudes during winter is a matter of record.

In a document distributed by the WMO [1], all phases of the balloon handling, storage, and conditioning problem were discussed. Besides the conventional methods of conditioning the sounding balloon (moist heat-box, immersion in boiling water), the use of petroleum products was suggested as a result of some experiments carried out in the Union of Soviet Socialist Republics. These tests [2] indicated an increase during the cold months of 5–8 km. in the bursting heights of balloons that were conditioned with liquid hydrocarbons.

Due to the advance planning that was necessary for the operation in Antarctica, supplies (including rawinsonde balloons) for some of the stations had to be shipped to Antarctica in 1955 and stored there until the establishment of the permanent bases in the summer of 1956–57. Storage of neoprene balloons at low temperatures (in this case 0° to −50° C.) results in a partial loss of elasticity. It was not at all surprising, therefore, that these balloons did not attain the expected heights and very frequently failed to reach the minimum of 50 mb. requested for the IGY.

2. EQUIPMENT

With two exceptions, all of the balloons used at the South Pole Station during the period reported here were 500-gram balloons (both day and night flight) dated 1955. The night flight balloons were used from the initiation of the program until September 25, the day flight balloons from September 12 to November 30. These balloons had been stored at McMurdo, Antarctica for nearly a year before they were airdropped to the South Pole Station. The balloons in use at the beginning of the program had been in warm storage (+20° to +25° C.) for about a month before use, and immediately prior to use the balloons were conditioned in a heat-box at +55° C. for 12 to 24 hours. For the moist heat conditioning, pans of water were placed on the floor of the heat-box and the balloons conditioned for 36 hours at a temperature of about +60° C.

A Gill low-pressure hydrogen generator was used for manufacturing the inflating gas. Approximately 20 to 30 minutes were required to inflate the balloon to a free lift of 1,900 to 2,300 grams. Whenever two or more charges were required to get a usable balloon, the second and succeeding charges produced a noticeable increase of

FIGURE 1.—Typical non-symmetrical balloon. This balloon was used for flight no. 238, 1200 GMT, July 29, 1957. It burst at 120 mb., 13,200 m. The largest diameter is about 8.5 ft.

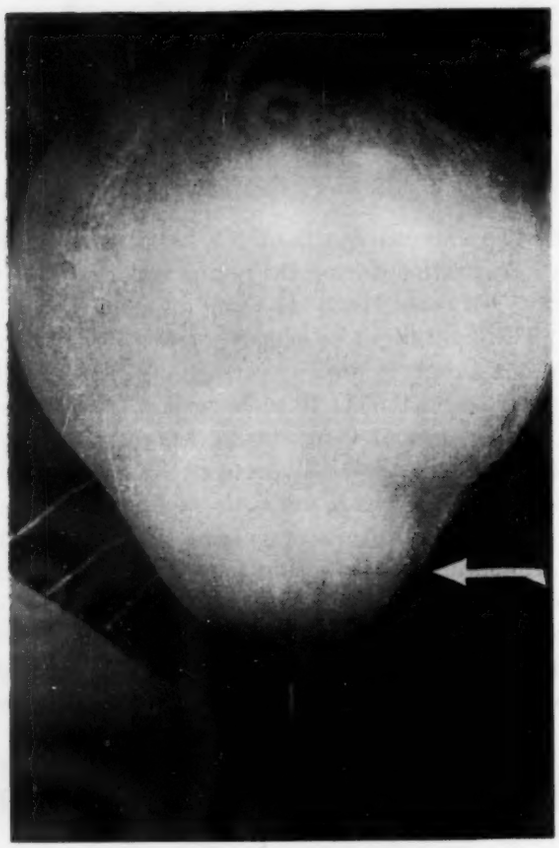

FIGURE 2.—Typical non-symmetrical balloon. This balloon was used for flight no. 329, 0000 GMT, September 13, 1957. It burst at 153 mb., 12,021 m. The arrow indicates the center line of the balloon.

steam along with the hydrogen, the steam subsequently condensing on the inner walls of the balloon. Depending on the temperature of the inflation room the condensed vapor was either water droplets or ice crystals. A gravity condenser was used in the gas line; for second or third inflations a second condenser can was used with both cans packed in snow. Nevertheless, many flights were made with water or ice crystals completely covering the inner walls of the balloon. It is worth noting that a large percentage of such flights reached average or better altitudes. During the period May through October, 56 flights were made with second and third inflation balloons with an average bursting level of 68.2 mb. This compares with an average bursting level of 64.1 mb. for single inflation balloons during the same period. For three of these months the second and third inflation balloons had higher termination heights than the single inflation balloons.

Throughout the year, better than half of the balloons inflated were non-symmetrical in shape. Figures 1 and 2 show typical examples. Generally this type of balloon developed a bulge which inflated at the expense of the rest of the balloon (note the position of the center line of the balloon in the second photograph). In nearly every case a well-shaped symmetrical balloon gave a high flight, while there was considerable variability in the performances of the non-symmetrical balloons.

3. FLIGHT TESTS

The first rawinsonde flight was made at the South Pole Station on March 27, 1957, at a time when the stratosphere was already appreciably cooled from its very warm summer state (compare figs. 3 and 8), although it was no colder than is normally found at a mid-latitude station. Despite this, the average sounding height for the first month was sub-par, averaging near 16.5 km., and dropping even lower in late April. Various conditioning methods were then tried. Immersion in gasoline had been suggested but this method was given only a short test because of the fire hazard. Instead, Arctic diesel fuel was tried as a conditioner. Soaking of the balloon in diesel fuel (or gasoline) caused swelling and increased the elasticity of the neoprene and immediately resulted in improved sounding heights.

TABLE 1.—*Results of soundings with balloons conditioned by three methods during trial period, May 1–7, 1957*

Number of flights	Conditioning method	Average pressure at burst (mb.)	Average height at burst (m.)	Average temperature at burst (° C.)	Pressure at burst	
					Best flight (mb.)	Worst flight (mb.)
6	Dry heat-box	125.5	14,038	−68.9	70	211
5	Diesel fuel	42.4	20,447	−74.8	23	74
2	Gasoline			−70.5	89	294

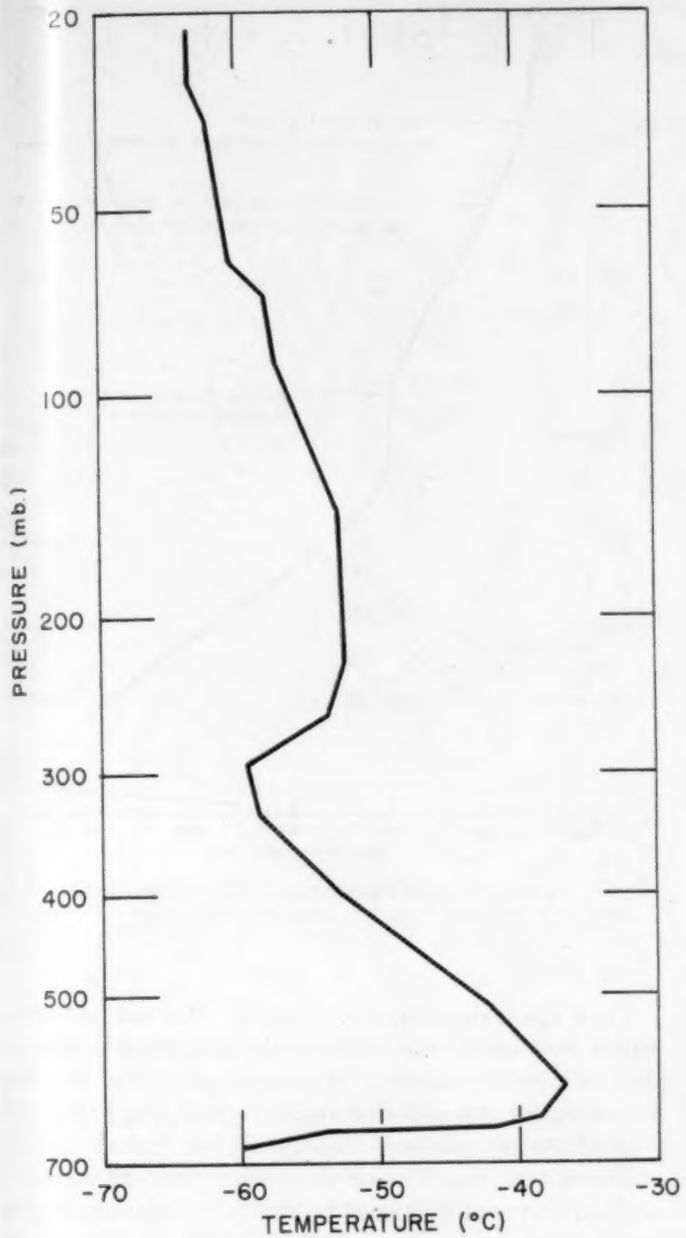

FIGURE 3.—Typical early fall sounding, 0000 GMT, April 4, 1957.

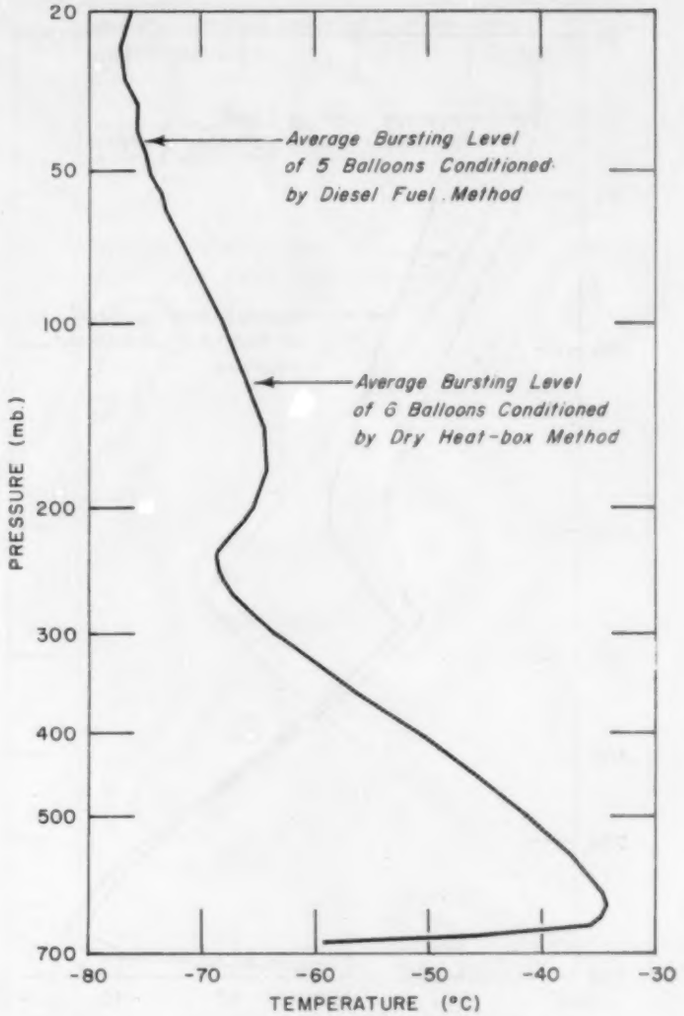

FIGURE 4.—Mean sounding for May 1–10, 1957.

On May 1, the first flight was made with a balloon conditioned by immersion in warm diesel fuel. For a period of a week following this initial flight successive soundings were made using balloons conditioned by three different methods. The results are summarized in table 1. In the following tables only flights terminated by bursts are considered.

Although the sample was small the results were impressive. Figure 4 is a plot of the mean sounding for the period May 1–10, with some of the information from the table.

To look a bit further into the suggestion that the diesel conditioning improved the sounding heights, 22 heat-box flights immediately preceding the above trial period were compared with the 22 diesel flights immediately after the trial period. The results are summarized in table 2.

These statistics are more impressive when it is realized that the stratosphere was colder for the diesel balloons than for the dry-heat balloons. This is illustrated in figure 5.

TABLE 2.—*Comparison of soundings with balloons conditioned by heat-box method (April 18-May 1, 1957) and by diesel fuel method (May 7–23, 1957)*

Number of flights	Conditioning method	Average pressure at burst (mb.)	Average height at burst (m.)	Average temperature at burst (° C.)	Pressure at burst	
					Best flight (mb.)	Worst flight (mb.)
22	Dry heat-box	86.6	16,205	−64.8	50	127
22	Diesel fuel	33.3	21,751	−75.7	11	59

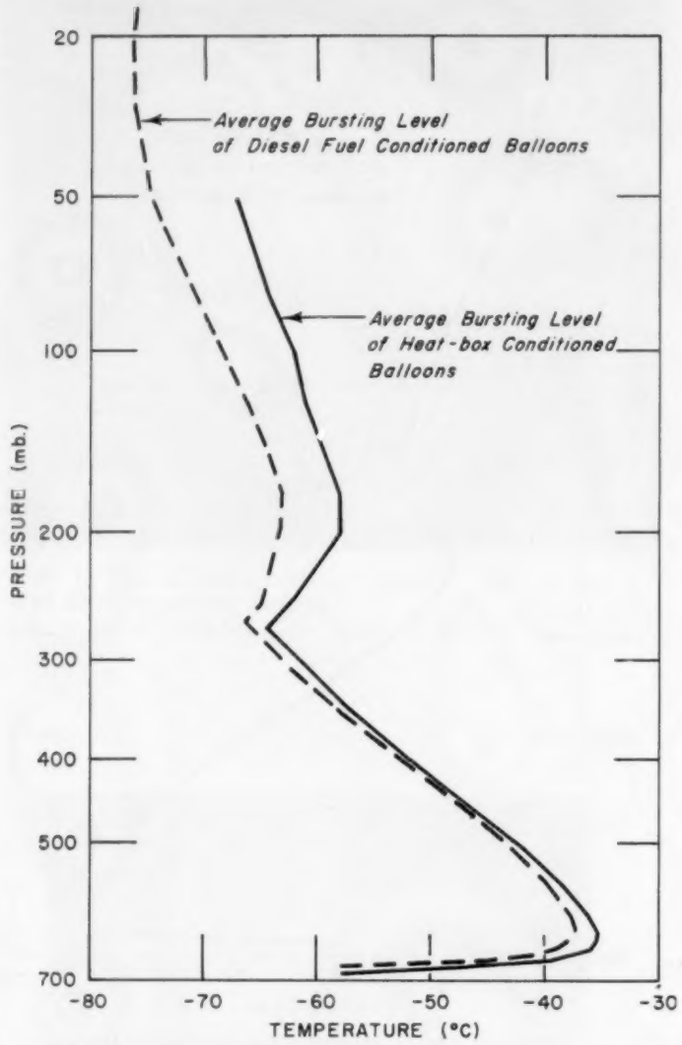

FIGURE 5.—Mean sounding with balloons conditioned by heat-box method, April 18–May 1, 1957 (solid line) and by diesel fuel method, May 7–23, 1957 (dashed line).

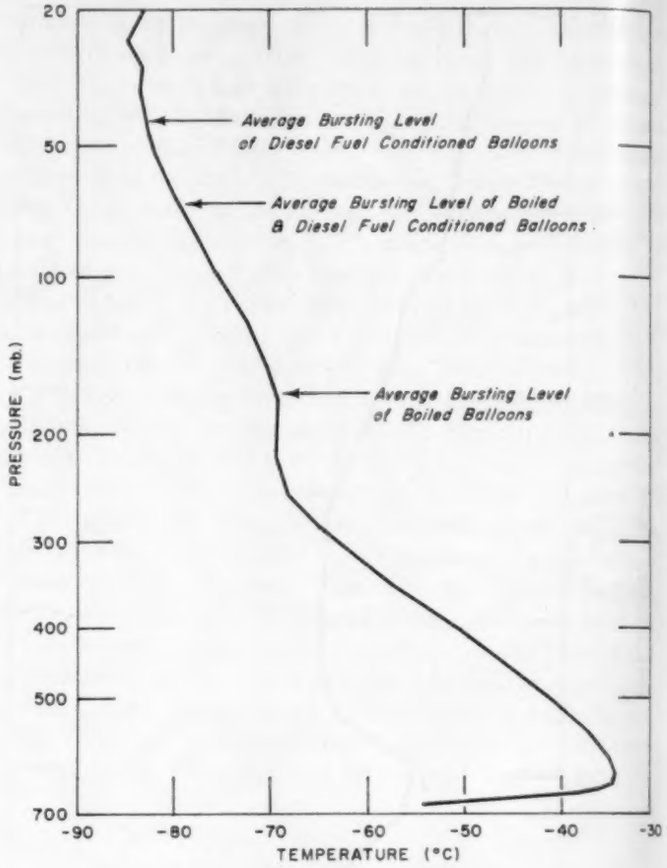

FIGURE 6.—Mean sounding for June 21–30, 1957.

In mid-June a series of flights was made to compare diesel fuel conditioning with the more conventional boiling water method of conditioning. The tests also included a third method—immersion of the balloon in boiling water followed by immersion in diesel fuel. Table 3 presents these results.

Once again the sample was small. The test was terminated because of the consistently poor flights obtained with the boiled balloons. Figure 6 is a plot of the mean sounding for the period June 21–30 indicating the condition of the stratosphere during this test period.

Diesel fuel conditioning was used throughout the winter and early spring with good results. During these months, a few flights were made with dry-heat conditioned balloons with poor results in every instance. Figure 7 is a plot of two soundings, one with a cold troposphere and the other with a cold stratosphere, typical of the atmosphere through which the balloons were ascending in the period July through September.

TABLE 3.—*Comparison of soundings with balloons conditioned by boiling water and by diesel fuel, June 17–30, 1957*

Number of flights	Conditioning method	Average pressure at burst (mb.)	Average height at burst (m.)	Average temperature at burst (° C.)	Pressure at burst	
					Best flight (mb.)	Worst flight (mb.)
4	Boiling water	169.0	11,742	−70.1	135	200
10	Boiling water and diesel fuel	69.1	17,588	−79.0	32	154
10	Diesel fuel	42.7	19,945	−82.7	19	77

TABLE 4.—*Comparison of soundings with balloons conditioned by three methods, November 1957*

Number of flights	Conditioning method	Average pressure at burst (mb.)	Average height at burst (m.)	Pressure at burst	
				Best flight (mb.)	Worst flight (mb.)
35	Diesel fuel	76.5	19,060	18	262
10	Boiling water	22.7	26,458	14	34
9	Moist heat-box	43.9	23,859	15	142

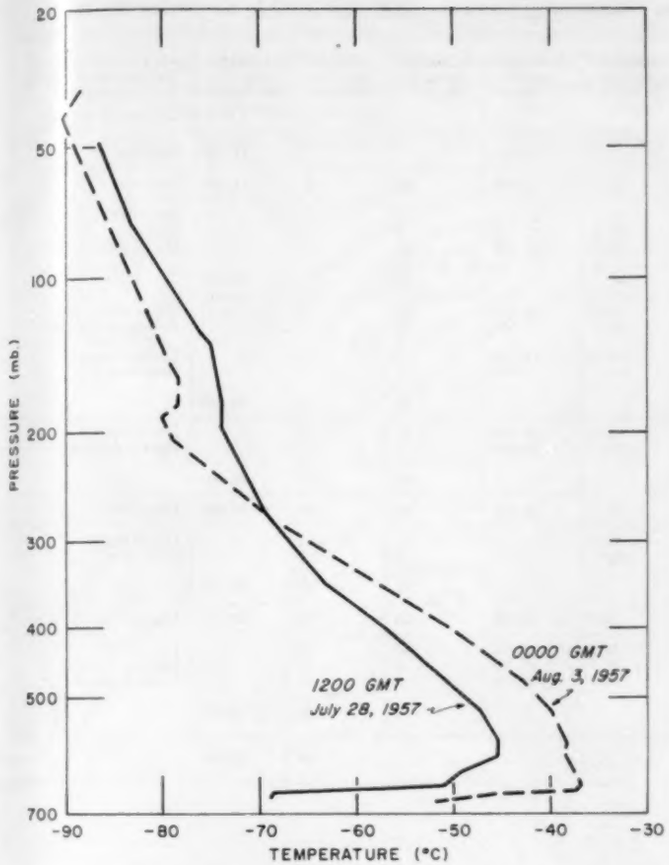

FIGURE 7.—Typical late winter soundings, 1200 GMT, July 28, 1957 (solid line) and 0000 GMT, August 3, 1957 (dashed line).

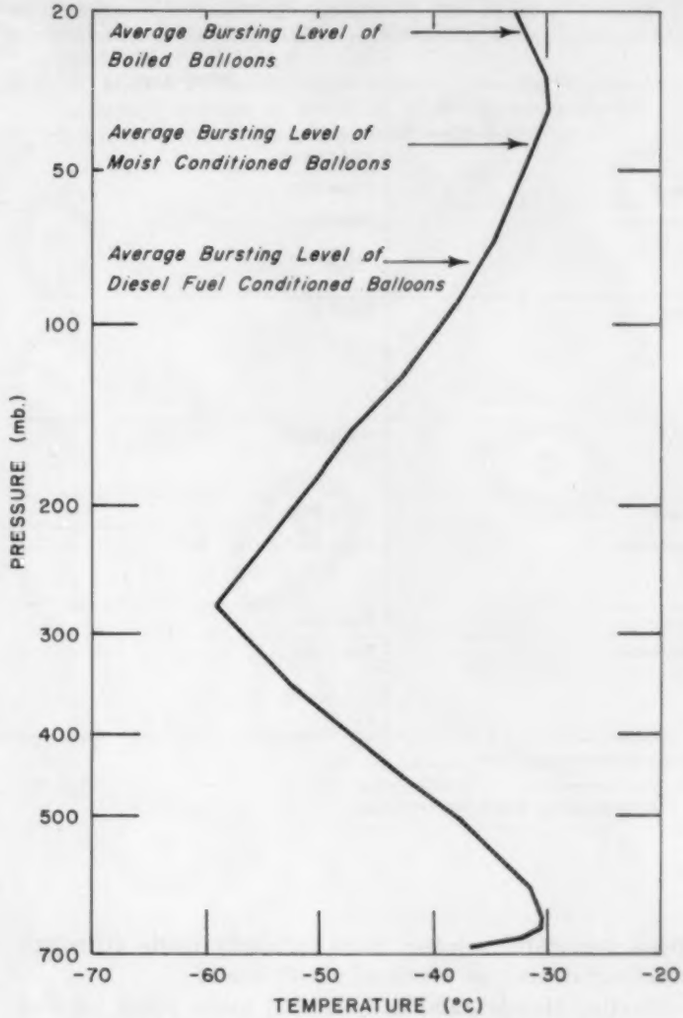
Average Bursting Level of Boiled Balloons
Average Bursting Level of Moist Conditioned Balloons
Average Bursting Level of Diesel Fuel Conditioned Balloons
PRESSURE (mb.)
TEMPERATURE (°C)

FIGURE 8.—Mean sounding for November 1957.

During the month of October the Antarctic stratosphere undergoes a marked warming, beginning and being most marked at the highest levels sounded. During October 1957 at the South Pole Station, the temperature at 20 mb. warmed from −67° C. on the 1st to −24° C. on the 30th, nearly 1.5° C. per day. With the warmer stratospheric temperatures the bursting heights of the balloons began to rise rapidly. However, during the first half of November there was an indication that the diesel-conditioned balloons were no longer producing consistently high flights. Significantly better flights were made at this time by using balloons conditioned by immersion in boiling water. Table 4 presents the results for November and includes both diesel conditioning and conditioning with boiling water along with some results from balloons treated in a moist heat-box (relative humidity 100 percent).

Figure 8 is the mean sounding for the month of November 1957. It can be seen that the stratosphere was quite warm; temperatures below −70° C. were seldom encountered.

Table 5 summarizes the results from all flights for the period March 27 to November 30, 1957, tabulated according to the different conditioning methods employed.

4. REMARKS

No tests were made to determine the optimum temperature of the diesel fuel in which the balloons were conditioned although this point did not appear critical. In practice the temperature of the diesel fuel ranged from about −10° C. to +40° C. Several checks were made of the effect of prolonged immersion in the diesel fuel and a few flights were made with good results using balloons that had been immersed for 24 hours. These balloons exhibited extreme swelling (the necks swelled to nearly twice the diameter of an untreated balloon) and were very fragile. From the limited evidence it did not appear that lengthy immersion (greater than ½ hour) improved the balloons' performance. Although the diesel-conditioned balloons were probably more fragile than dry bal-

TABLE 5.—*Comparison of all soundings classified according to balloon conditioning methods used, March 27 to November 30, 1957*

Month	Balloon type	Average pressure at burst (mb.)	Average height at burst (m.)	Number of burst flights	Total number of flights	Average height of all flights (m.)	Balloon conditioning method
March	Night flight		16,005	4	5	14,091	Heat-box.
April	Night flight	83.4	16,652	38	46	15,475	Heat-box.
May	Night flight	125.7	13,969	7	7		Heat-box.
		161.5	12,622	2	2		Gasoline.
		36.5	21,268	29	42		Diesel fuel.
				38	51	18,652	
June	Night flight	42.6	20,158	40	56		Diesel fuel.
		69.1	17,588	10	10		Boiling water and diesel fuel.
		169.0	11,742	4	4		Boiling water.
		333	7,637	1	1		Heat-box.
				55	*71	18,243	
July	Night flight	84.0	16,400	65	70		Diesel fuel.
		218	10,025	1	1		Lubricating oil.
				66	*71	16,102	
August	Night flight	84.6	15,885	60	62	16,010	Diesel fuel.
September	Night flight and day flight	73.0	17,123	56	61		Diesel fuel.
		340	7,072	1	1		Heat-box.
				57	*62	16,601	
October	Day flight	44.7	20,208	58	61	20,034	Diesel fuel.
November	Day flight	76.5	19,060	35	36		Diesel fuel.
		22.7	26,458	10	12		Boiling water.
		43.9	23,859	9	11		Moist heat-box.
				54	59	20,603	
March 27–November 30, 1957					488	17,688	

*Includes special World Day soundings.

loons, successful releases were regularly made (through overhead doors) in winds of 20–30 knots.

Martin, Mandel, and Stiehler [3] made patch tests of neoprene balloons and concluded that the two important properties in determining the flight elevations of the balloons were the flaccid length (uninflated length) of the balloon and the elongation of the neoprene film (ratio of the final to initial length of a line drawn on the film). Their equation for predicting the flight elevation is:

$$h = 43.8 \log (E_B L_O) - 86.3$$

where h is the bursting elevation in kilometers, E_B is the elongation of the film, and L_O is the flaccid length of the balloon in inches. This equation used with the data from table 2 of the present paper indicates an increase of about 30 percent in the elongation of the neoprene film of the diesel-treated over the untreated balloons. The effectiveness of various conditioning methods could be checked in the laboratory by using the techniques developed by Martin, Mandel, and Stiehler.

5. CONCLUSIONS

Although the test samples reported in this paper were necessarily small, certain features appear sufficiently consistent to warrant the following conclusions:

1. Soaking the sounding balloon (both inside and out) in diesel fuel resulted in definitely higher flights during the dark, cold months, and possibly will be effective in the Antarctic and Arctic [1] when upper-air temperatures are colder than −70° C.

2. During the cold months, boiling water or dry-heat conditioning of the balloon were inferior conditioning methods.

3. However, with an atmosphere (above the surface inversion) where the coldest temperatures were −70° C. or higher, conditioning of the balloon by immersion in boiling water [2] or by exposure in a moist heat-box were definitely superior to diesel fuel conditioning.

4. Although the conditioning process is very important, the quality and condition (considering storing and handling) of the balloon are of equal importance.

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[1] According to the Instrumental Engineering Division, U.S. Weather Bureau, there is some recent evidence from limited tests in the Arctic which appears to show that balloons that have had a greater amount of post plasticizing than those used at the South Pole Station do not respond to the diesel fuel conditioning.

[2] Also according to the Instrumental Engineering Division, there is some recent evidence from limited tests which indicates that balloons that have been heavily post plasticized by the manufacturer react unfavorably to the boiling water method of conditioning.

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MINIMUM TEMPERATURES IN THE LOWER STRATOSPHERE

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[Manuscript received November 28, 1958; revised July 24, 1959]

ABSTRACT

An analysis of the lowest temperatures observed at the base of the stratosphere in various zones of the globe shows that the recent records obtained over the Antarctic lower the previous world record of long standing, which was registered in the equatorial zone at Batavia in 1913.

1. INTRODUCTION

The thermal regime of the stratosphere is now the object of a thorough investigation. The steady flow of new temperature data obtained in the upper troposphere and lower stratosphere under the International Geophysical Year program enhances our knowledge, but also raises new problems in our understanding of the global stratosphere. The new data need critical evaluation. This, in turn, calls for a survey and for an analysis of the earlier records, to serve as a basis for comparisons with the later ones.

In this article, the lowest observed temperatures of the lower stratosphere are surveyed and analyzed on the global scale, to provide a background for further investigations in this field.

2. BATAVIAN RECORD

The first record low temperature in the lower stratosphere was registered by a sounding balloon over Batavia (Java) on November 5, 1913. It was −91.9° C. (−133.4° F.) at a height of 17,330 m.

Until 1958 this represented the lowest limit of temperature for the base of the stratosphere. Although in this ascent the clockwork had stopped, the register was accepted by W. van Bemmelen [3,4] who published it, without any reservation. A complete record of the first sounding, as van Bemmelen [3] gave it, is reproduced in table 1.

TABLE 1.—*Registering balloon ascent, Batavia, November 5, 1913*

Local time (hr.)	(min.)	Altitude (m.)	Pressure (mm.)	Temperature (°C.)	Remarks
4	25	8	750	23.6	
				25.7	Clock stops. Inversion.
	0.00	3,950	477	3.3	Clock starts again.
		4,000			
	1.43	4,320	456	2.1	
		4,500			
	2.78	4,770	431	−0.9	
		5,000			
	5.50	5,440	396	−6.0	
		5,500			
		17,330	65	−91.9	Maximum height, minimum temperature.
		10,720	198	−43.0	
		8,740	258	−28.0	
		6,160	364	−9.1	
		25	760	27.1	

Since this record represents a rather isolated reading, its authenticity would be questionable if it were not supported by another, very detailed and high-reaching sounding, made at the same location a month later. The second sounding during which very low temperatures were recorded at Batavia was conducted on December 4, 1913. The registering balloon reached 26,040 m. and the readings are to be depended upon. The lowest temperatures registered during this ascent were: −89.5° C (−129.1° F.) at 17,000 m. on the way up, and −99.9° C. (−131.6° F.) at 16,500 m. on the way down. There is a very complete tabular record of this sounding [3] which is reproduced here in graphic form in figure 1.

To complete the discussion of the Batavian world record of lowest temperature at the base of the stratosphere, it should be mentioned that only one case of a competing value (before the recent Antarctic records) was found in meteorological publications. This however, as we shall show, is disqualified by proper analysis of the sounding. A. Wagner [24] mentioned that a minimum temperature of the upper air of −92° C. was registered in Agra (India) (27°10′ N., 78°05′ E.), which was equal to that recorded over Batavia. Wagner derived this information from G. Chatterjee and N. K. Sur [5], who, in turn, had reproduced a graph from K. R. Ramanathan [17]. In reproducing it they made an error, showing one dot too many and in a wrong place. It was this dot that showed the extremely low temperature of −92° C., which had neither been shown nor discussed in the original publication of the Agra upper-air temperatures by Ramanathan

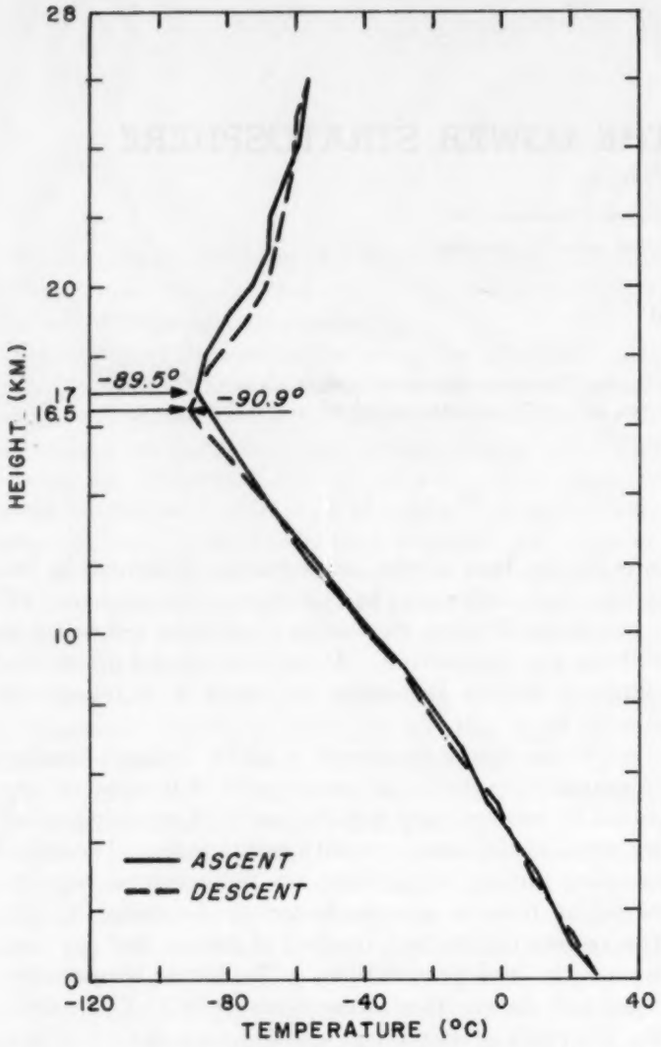

FIGURE 1.—Upper-air sounding at Batavia, Java, December 4, 1913.

[16]. Also incorrect in the same source [5] is the year of the Batavian records, which are ascribed to December 5 and November 4, 1923. These records were registered in the year 1913, as the original paper by van Bemmelen [3] indicates.

3. LATER LOW-LATITUDE DATA

It is remarkable that since the time of the famous Batavian record in 1913 no lower temperatures were observed in the stratosphere until recently. To illustrate the usual values of the lowest temperatures obtained in the atmosphere over the low latitudes, data extracted from [20] for three stations in the Caribbean region are given in table 2. These data indicate, however, that temperatures close to the Batavian records probably could be obtained in this region if the soundings extended just one kilometer higher; i.e., reaching the levels of the Batavian records, which were registered at 17,000 and 16,500 meters.

TABLE 2.—*Lowest temperatures (°C.) by height for three stations in Caribbean region (from [20])*

Height (km.)	San Juan, P.R., 1941–45	Swan Island, 1940–45	Tacubaya, Mexico, 1943–45
Surface	17	18	4
1	13	12	
2	7	6	
3	2	3	3
4	−3	−3	−2
5	−9	−9	−9
6	−15	−17	−15
7	−22	−20	−22
8	−29	−27	−28
9	−36	−35	−37
10	−46	−43	−46
11	−55	−51	−54
12	−62	−59	−62
13	−69	−68	−70
14	−75	−76	−73
15	−82	−82	−70
16	−88	−87	−85

D. Dewar [7], in his recent brief communication, has mentioned an isolated value of −97°C. recorded at 100 mb. in Verval (India), and −94°C. at the same level in Columbia (Ceylon) during 1944–50.

Observations on minimum temperatures of the tropopause in Bangui, French Africa (4°22′ N., 18°34′ E.; H=386 m.) during the period 1953–57, as given in a recent publication [12], showed the following lowest temperatures by months:

	°C.		°C.
January	−83.9	July	−81.4
February	−85.6	August	−80.0
March	−88.0	September	−82.0
April	−85.8	October	−84.7
May	−86.4	November	−85.5
June	−81.5	December	−84.4

The absolute minimum was observed in March, but the publication does not say at what altitude this temperature was registered.

4. MIDDLE-LATITUDE DATA

It is well known that in the middle latitudes the base of the stratosphere is warmer and temperatures there

TABLE 3.—*Lowest temperatures (°C.) by height for four stations in the middle latitudes (from [20]). The lowest minima are underscored*

Height (km.)	El Paso, Tex. (31°48′N.), 1939–45	Albuquerque, N. Mex. (35°05′N.), 1943–45	Denver, Colo. (39°45′N.), 1939–45	Rapid City, S. Dak. (44°02′N.), 1943–45	Glasgow, Mont. (48°11′N.), 1943–45
Surface	−17	−27	−34	−37	−57
1				−24	−26
2	−11	−12	−22	−24	−25
3	−13	−17	−26	−26	−28
4	−19	−23	−33	−33	−34
5	−26	−30	−37	−40	−41
6	−33	−38	−41	−45	−46
7	−41	−46	−49	−52	−54
8	−49	−50	−53	−55	−61
9	−53	−55	−58	−59	−62
10	−57	−60	−65	−64	−63
11	−64	−65	−68	−66	−67
12	−68	−69	−72	−70	−71
13	−72	−70	−71	−71	−72
14	−79	−71	−72	−68	−71
15	−77	−76	−73	−70	−72
16	−80	−77	−76	−70	−66

cannot compare with the records from the equatorial or tropical zone. Table 3 gives data for several stations in the United States, approximately along the 105th meridian, listed in the order of increasing latitude. More recently a temperature of −80° C. at 33 mb. was recorded at Salem, Oreg., on February 4, 1957 [22].

5. HIGH-LATITUDE DATA

ALASKA

The pattern of the vertical distribution of the lowest temperatures over Alaska is shown in figure 2a (from [20]). The lowest temperatures obtained by radiosoundings are: In Fairbanks −72° C. (−97.6° F.), 1939–45, at 11 km.; in Nome −69° C. (−92.2° F.), 1940–42, at 12 km.; in Barrow −74° C. (−101.2° F.), 1940–45, at 11 km. For comparison, data for two low-latitude stations (from [20]) are also displayed in figure 2b.

GREENLAND

The latest published data for Greenland come from the expedition led by Paul-Emile Victor from September 1949 to August 1951. During this period 192 soundings were made at the French station (70°55′ N., 40°38′ W.; H=2993 m.). The lowest temperature was recorded at approximately the 150-mb. surface on March 11, 1950, at 3 p.m. It was −74° C. (−101.2° F.). This ascent (from [1]) is shown in figure 3.

Wexler and Moreland [25] mention "stratospheric temperatures found at Thule, Greenland, in late January 1952 . . . as low as −81° C".

ARCTIC REGIONS

Radiosoundings made from the Soviet ice floe station "North Pole–4" during the period from April 1955 to April 1956 obtained, for the first time, more or less complete data on air temperatures aloft over the Central Arctic region [11]. The ascents reached up to 35,540 m.

In October and December the temperature in the lower stratosphere usually decreased with height, whereas the troposphere was still of an inversional or isothermal type. After November 30, a decrease in temperature with height was observed in the troposphere also. Thus, for the midwinter, a troposphere with a slightly falling temperature, and a more rapid fall of temperature in the lower stratosphere, was typical.

In December 1955, at an altitude of 18.5 km., a temperature as low as −77.4° C. (−107.3° F.) was recorded for the first time. It was observed in a high-level cyclone which had its center over Baffin Bay. In the upper troposphere, at heights of 10–11 km., strong westerlies of more than 75 m./sec. indicated the existence of great thermal contrasts and pressure gradients, as effected by an intensely cold air mass. On December 6 and 7, temperatures of −78° C. (−108.4° F.) and −79° C. (−110.2° F.) were recorded in the lower stratosphere, and on January 4, 1956, the lowest temperature on record for this region was observed, −81.2° C. (−114.2° F.). It was registered at an altitude of about 20 km. At that time the drifting station was in the region of 85° N., 179° E. This ascent is reproduced from the source [11] in figure 4. The atmospheric processes in 1955–56 were rather abnormal, especially so in the midwinter when an extremely strong cyclonic activity developed; this could be one of the causes for a colder stratosphere.

The records of lowest temperatures obtained by earlier drifts in 1953–55 by "North Pole-3" and "North Pole-4" gave −79.2° C. (−94.4° F.) at 8.9 km. on December 21,

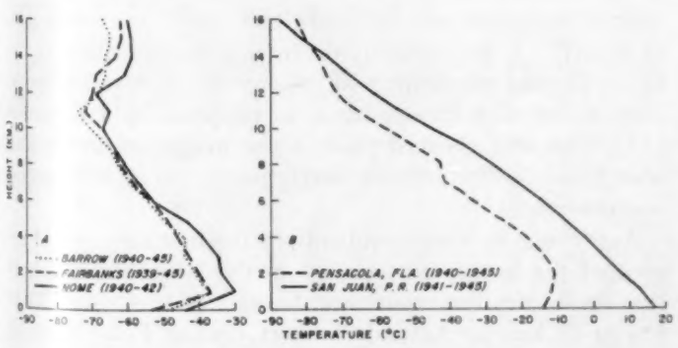

FIGURE 2.—Vertical distribution of lowest temperatures at (a) three Alaskan stations, and (b) two low-latitude stations.

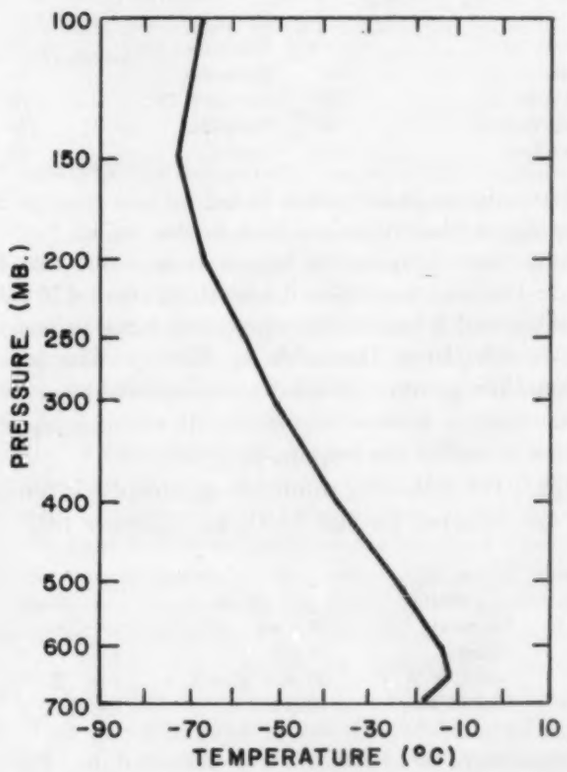

FIGURE 3.—Upper-air sounding at French station in Greenland, March 11, 1950 (from [1]).

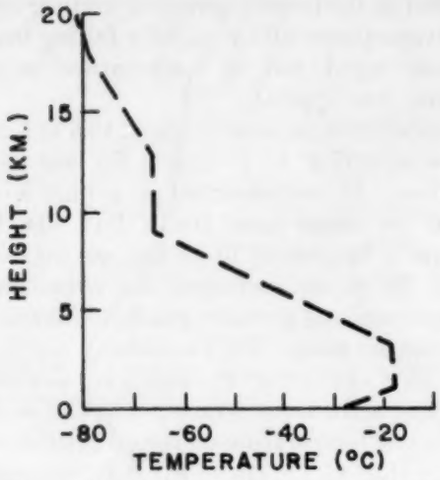

FIGURE 4.—Upper-air sounding at drifting station "North Pole-4" (85° N., 179° E.), January 4, 1956.

1954 and −71.6° C. (−96.9° F.) at 9.5 km. on November 13, 1954, respectively. At those times the soundings were limited to 12–14 km.

Belmont [2] gives the following data on minimum temperatures of the stratosphere as recorded from the Ice Island T–3. This island was floating in the Central Arctic during the period June 20, 1952–March 4, 1953. The absolute minimum temperatures derived from minimum curves are as follows:

	°C.		°C.
June 1952	(−58)	November 1952	−57
July	−57	December	−63
August	−58	January 1953	−68
September	−62	February	−59
October	(−62)	March	(−62)

The parentheses show values based on less than 10 days data. The author indicates that in this region ". . . the apparent mean tropopause height is near 300 mb. from June to October, and then it lowers to about 370 mb. in November and it may be interpreted as lying between 350 and 370 mb. from December to March. The level of minimum temperature generally occurs about 25 or 50 mb. higher than the level of significant decrease in lapse rate at which we define our tropopause."

In [22] the following minimum stratospheric temperatures are reported for the Arctic for the year 1957:

Temperature	*Station*	*Height*	*Date*
−81° C.	Ice Skate "A"	19.6 km. (45 mb.)	Dec. 21
−82	Eureka	25 mb.	Dec. 11, 26, 30
−81	Alert, N.W.T.	30 and 40 mb.	Jan. 23
−80	Mould Bay	40 mb.	Jan. 29

A temperature of −83° C. was recorded by Bjerknes (1939) in As, 60 km. south of Oslo, at 23.5 geodynamical km. as published by H. Dieterichs [8].

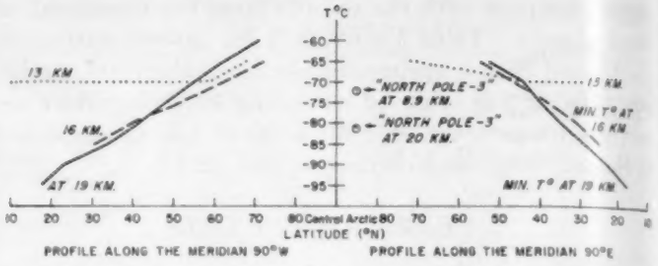

FIGURE 5.—Profiles of minimum temperatures at three levels in the Northern Hemisphere.

6. NORTHERN HEMISPHERE

To consolidate the knowledge of the lowest temperatures in the upper layer over the Northern Hemisphere the distribution of the lowest temperatures observed at the levels of 13, 16, and 19 km. is presented on a profile graph along the 90th meridian in figure 5. This graph is based on data given in [21] and on observations made by drifting stations "North Pole-3" and "North Pole-4". Figure 5 affords a general view of the variations of the lowest temperatures with latitude at three altitude levels. Since the maps of [21] present some interpolated values which have never been observed but only theoretically computed, especially for the 19-km. level, so does the graph.

7. ANTARCTICA

Paradoxical as it may seem, it was over the Antarctic continent that the temperatures competing with the Batavian world record were registered, for the first time at Maudheim (−91° C.) in the winter of 1950 [18], and later at Amundsen-Scott (South Pole) station (−91.2° C.) in the winter of 1957 [10].

The records of radiosoundings in Antarctica have frequently indicated the fact that, in the midwinter season, the temperature diminished with height continuously throughout the troposphere and in the lower stratosphere as well. The tropopause was frequently missing. This has led to the conclusion that, in the middle of the winter season, the tropopause in the Antarctic actually disappears. Such a statement was first made by Court [6], and later it was confirmed by Rusin (see [14]), the aerologist at Mirny. A continuously decreasing temperature, up to 16 or 17 km., reaching −75° and −76° C. was recorded over Mirny and Pionerskaya, as reported by Laktionov [14], who also claimed that at the height of 18–20 km. over Mirny, a distinctive westerly jet stream of 100 m./sec. was observed.

According to recent publications the first new world record of the lowest temperature in the lower stratosphere that broke the Batavian record was −93.0° C. (−135.4° F.) at 21 km. at Amundsen-Scott (South Pole) station, 0000 GMT, July 17, 1958, as reported by the U. S. National Academy of Sciences [15].

TABLE 4.—*Radiosonde ascents which recorded the lowest temperatures in the upper layers in 1958 in Antarctica*

Pressure (mb.)	Amundsen-Scott (South Pole) (90° S., H=2800 m.), July 17, 0000 GMT	Byrd (79°59′ S., 120°01′ W.; H=1515 m.), July 15, 1200 GMT	D'Urville (66°40′ S., 140°01′ E.; H=40 m.), July 28, 0000 GMT	Hallett (72°18′ S., 170°18′ E.; H=5 m.), Sept. 4, 1200 GMT	Little America (78°11′ S., 162°10′ W.; H=45 m.), July 10, 1200 GMT	McMurdo (77°51′ S., 166°37′ E.; H=45 m.), Aug. 28, 0000 GMT	Mirny (66°16′ S., 110°31′ E.; H=30 m.), July 20, 1200 GMT	Sovetskaia (70°24′ S., 87°35′ E.; H=3570 m.), Aug. 8, 0000 GMT	Wilkes (66°35′ S., 110°31′ E.; H=12 m.), Aug. 28, 0000 GMT
	°C.	°C.	°C.	°C.	°C.	°C.	°C.	°C.	°C.
24					−91				
25	−93	−92							
27					−91				
30	−93								
32									−81
37						−87			
44				−84					
45								−91	
47	−89		−73						
50	−91	−89	−72	−84	−89	−85		−91	−78
65							−84		
100	−83	−83	−68	−76	−80	−81	−72	−86	−73
150	−76	−77	−66	−72	−75	−75	−68	−78	−66
200	−72	−75	−66	−68	−73	−73	−65	−76	−60
300	−65	−67	−61	−60	−65	−66	−58	−70	−52
400	−53	−51	−52	−49	−54	−56	−49	−62	−45
500	−43	−37	−44	−39	−43	−44	−41	−54	−41
550								−54	
575								−60	
591								−80	
597	−37								
640	−40								
666	−53								
700		−22	−25	−25	−26	−30	−28		−23
800				−33	−21	−33			
850							−20		−13
900				−31					
940									−9
980					−29	−29	−22		
995					−39	−29	−22		
1,000					−39				−12

The record of this radiosounding is reproduced in table 4 which shows also the records of other radiosoundings which registered the lowest temperatures in the upper layer at their respective stations in 1958. This table was compiled from the abundant aerological data collected and worked up by the Polar Meteorological Research Unit, U.S. Weather Bureau, which were made available to the author by the kind cooperation of the Chief of this Unit, Mr. M. J. Rubin, and his assistant, Mr. E. C. Flowers.

With the arrival of reports of the radiosoundings made on Falkland Islands [9] it became obvious that the lowest temperatures in the lower stratospheric layers had been registered by the Halley Bay station. In table 5 several records of radiosoundings at this station are reproduced showing temperatures in the lower stratosphere as low as −94°, −95°, −101°, and even −109° C. It seems to us that the readings of −109° and −101° C., as recorded on the 12th and the 17th of July at Halley Bay, are somewhat too low and the respective lapse rates too large to be trusted without further verification, but the temperature of −94° C., as observed on July 3, 11, and 27, and even the temperature of −95° C. observed on July 6, 1958 at the 18-mb. level seem to be reliable enough to be taken into consideration.

TABLE 5.—*Radiosonde ascents at Halley Bay, 75°31′ S., 26°36′ W., H=30 m., July 1958*

July 3 1200 GMT		July 6 1200 GMT		July 7 1200 GMT		July 11 1200 GMT		July 12 1200 GMT		July 17 0000 GMT		July 27 1200 GMT	
mb.	°C.	mb.	°C.	mb.	°C.	mb.	°C.	mb.	°C.	mb.	°C.	mb.	°C.
25	−94	18	−95	20	−93	20	−94	23	−109	28	−101	30	−93
37	−91	61	−90	26	−93	35	−90	35	−91	35	−93	52	−94
77	−85	111	−81	40	−88	45	−90	40	−87	44	−90	62	−91
242	−70	117	−79	50	−89	56	−86	67	−86	90	−84	72	−93
332	−60	164	−80	98	−80	68	−85	76	−84	118	−80	100	−86
420	−50	179	−78	149	−73	168	−75	85	−84	294	−79	104	−87
509	−40	608	−27	255	−68	190	−72	108	−79	407	−49	110	−85
604	−33	804	−15	315	−60	242	−73	125	−79	464	−44	135	−87
672	−28	999	−33	802	−16	282	−69	165	−73	648	−30	148	−85
827	−22			983	−23	312	−71	196	−76	760	−27	181	−85
1006	−32					661	−34	279	−68	840	−23	288	−60
						987	−46	295	−67	990	−16	400	−44
								625	−34			637	−23
								730	−28			988	−24
								894	−21				
								984	−28				

8. CONCLUDING REMARKS

Comparing the minimum temperatures in the lower stratosphere as recorded by several stations, and analyzing their reliability, one should consider the order of magnitude of those differences in temperature which are observed between individual radiosondes of various countries under equal conditions. Results of the comparison of temperature readings obtained by individual simultaneous soundings have been reported by Harmantas [13]. Using the data of simultaneous soundings made in Payerne, Switzerland, in May-June 1956, Harmantas compared the records of temperature obtained by radiosondes of 13 countries with those of the United States. It has been found that at surfaces higher than the 30-mb. level serious radiation errors exist. At night, the differences in temperature at the 50-mb. level varied within limits of −2° and +1.9° C., while the daytime soundings showed a much wider range of differences, limited by −18.1° and +5.6° C. Teweles and Finger [19] recently derived an empirical system for the reduction of diurnal variations in the reported temperatures and heights of stratospheric constant pressure surfaces from soundings made during the IGY.

At the present time it is not possible to ascertain that any one of the above-mentioned temperatures is the new world record of the lowest temperature for the lower stratosphere, and all that this article attempts to do is to provide a summary of the available data on this subject to facilitate future research.

ACKNOWLEDGMENTS

The author wishes to express her gratitude to Mr. Sidney Teweles, who suggested this investigation, to Mr. Morton J. Rubin and Mr. Edwin C. Flowers for valuable data, and to Mr. Benjamin Ratner and Mr. Roderick S. Quiroz for valuable advice.

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THE WEATHER AND CIRCULATION OF JULY 1959

The Second Consecutive July with an Unusual Circulation Pattern

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1. INTRODUCTION

For the second successive July [2] many of the troughs and ridges on the 700-mb. 30-day mean map (fig. 1) were not in the climatologically preferred locations indicated in studies made by Klein and Winston [6]. This was especially true of the western sectors of the Northern Hemisphere in July 1958, and again this July an abnormal trough appeared over the central United States. In the eastern and western coastal areas, where weak ridges were observed, the circulation was more anticyclonic than usual. This regime was associated with generally light precipitation in the West but heavy rainfall in several large areas east of the Rocky Mountain States (figs. 2, 3). Rainfall from tropical cyclones Cindy and Debra augmented the total precipitation in Texas and along the east coast. Temperatures averaged below normal in most of the Central and Southeastern States and were above normal in most other regions, with record heat reported in the Far Southwest (fig. 4).

2. 30-DAY AND 5-DAY MEAN CIRCULATIONS

The July circulation at the 700-mb. level over the western Northern Hemisphere (fig. 1) was one of small amplitude, and many of the troughs and ridges were out of phase with those on the July normal chart [9]. The normal July troughs along both the east and west coasts of North America were apparent this year only in eastern Canada and possibly along the California coast, where a trough remnant was discernible. More interesting is the unusual circulation in other areas. The ridges in the western Bering Sea, western United States, and north-central Atlantic, and the troughs in north-central Pacific, central United States, and northern Atlantic were all in climatologically infrequent areas [6]. The data of [6] show that 30-day mean ridges were extremely rare in the western Bering Sea in the Julys from 1933 to 1955, and in the northern latitudes over eastern Siberia no ridges at all were observed during this period. The 30-day mean ridge this July resulted from frequent incidence of 5-day mean ridges in this area (fig. 5). If the ridge occurrences at 60° N. for two 10° longitude bands are combined, we find that 5-day mean ridges were observed in July 1959 80 percent of the time between 155° E. and 175° E., compared with only 23 percent during the Julys of 1947–55.

The other aforementioned 30-day mean troughs and ridges were unusually placed to a similar degree. Note that the 30-day mean trough line in the central Pacific fell in a region of minimum trough frequency and in western Alaska passed through an area of zero trough frequency [6]. This July 5-day mean troughs were located at 60° N. in the 30° longitude band between 175° W. and 145° W. 100 percent of the time, compared to only 31 percent during July 1947–55. The 30-day mean ridge and trough in the United States each was located in an only slightly more favored area. The common situation in July is for a continental anticyclone to persist over central United States, but this July a 5-day mean trough was located at 40° N. between 105° W. and 85° W. 78 percent of the time (to be compared with 20 percent during the 1947–55 period).

It should be noted that the 30-day mean troughs and ridges (fig. 1) were located very close to the regions of high incidence of 5-day mean troughs and ridges, with minor exceptions in Asia (fig. 5), implying a good deal of stability in the general circulation during the month.

However, the mean circulation (fig. 1) differed in several aspects from that of June 1959 [4]. During June an exaggerated summertime pattern, with a ridge over central United States and a trough along each coast, was established. That this trend did not continue into July is attested by the anomalous height falls in central United States, flanked by rises of more than 200 ft. off the New England and Washington coasts (fig. 7). Although it is difficult to determine how or where these changes originated, they were accompanied by a train of height change centers extending upstream in the westerlies to about 120° E. In July 1958 the trough in central United States was likewise associated with abnormally located troughs and ridges as far upstream as Siberia [2].

Figure 8 is the 700-mb. 15-day mean height change from the first half to the last half of July 1959. During July there was considerable filling of the Arctic Low in the Beaufort Sea. In middle latitudes over the Pacific the perturbations intensified as falls occurred in the Siberian and Pacific troughs concurrent with rises in the adjacent ridges. Some retrogression of the western North American ridge is indicated by the large rise center, at 140° W., west of the ridge. In North America the trough

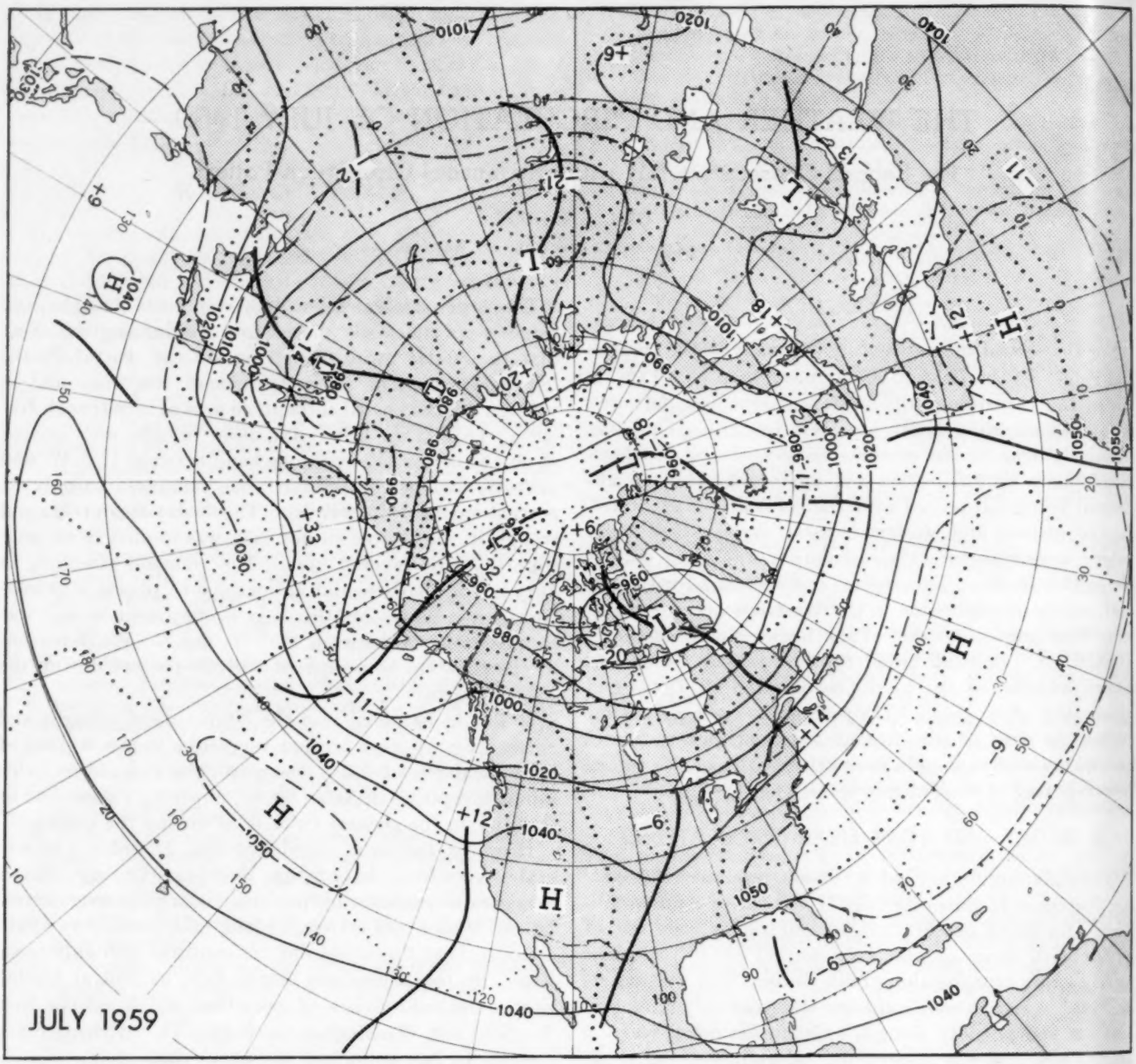


FIGURE 1.—Mean 700-mb. contours (solid) and height departures from normal (dotted), both in tens of feet, for July 1959. Many of the troughs and ridges were located in climatologically unfavored areas.

filled in the north but intensified over the Southern Plains States as July progressed. Over Europe an incipient block is discernible.

The 700-mb. 30-day mean jet stream (fig. 9) was rather ill-defined and displaced north of its normal position in many longitudes. This northward shift produced above normal wind speeds over Canada, but in the United States wind speeds were near to below normal. Negative departures along the east coast and in the western Atlantic (fig. 9B) reflect the easterly regime which prevailed in that area. Speeds were also well below normal in the western Pacific.

The largest monthly height anomalies, a +330 ft. and −320 ft. couplet, were associated with the Bering Sea ridge and a deep Arctic Low over the Beaufort Sea (fig. 1). This configuration produced a super-normal northwesterly flow over the Bering Strait, which brought cool thicknesses (1000–700 mb.) to much of Alaska (fig. 6), quite similar to March 1959 [3] when many places in Alaska reported record cold. Over the United States

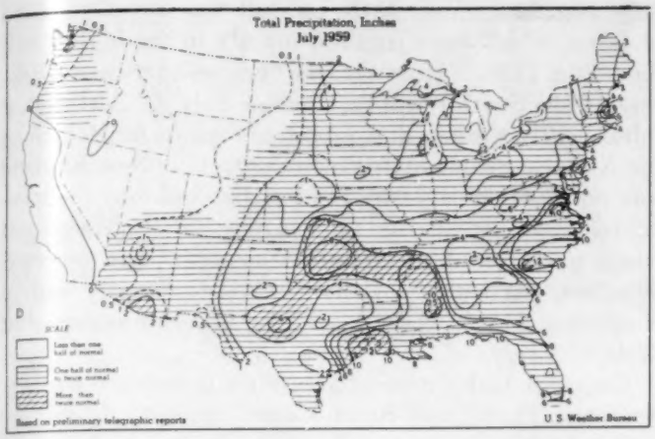


FIGURE 2.—Total precipitation (inches) for July 1959. Bulk of rainfall was east of the 30-day mean trough. (From [10] Aug. 3.)

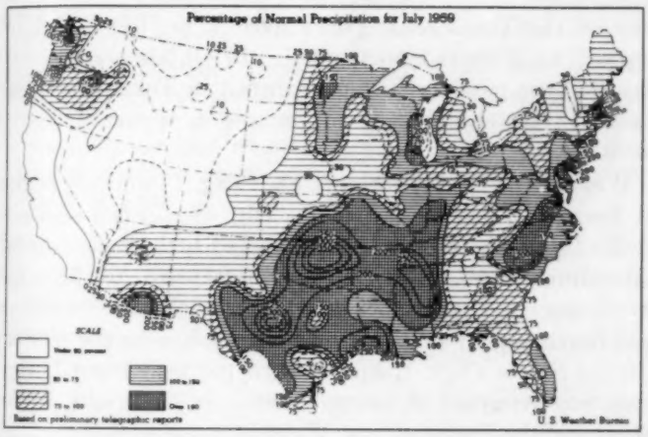


FIGURE 3.—Percentage of normal precipitation for July 1959. Rainfall was abnormally heavy in the South-Central and Middle Atlantic States but light in the northern Rocky Mountains. (From [10] Aug. 10.)

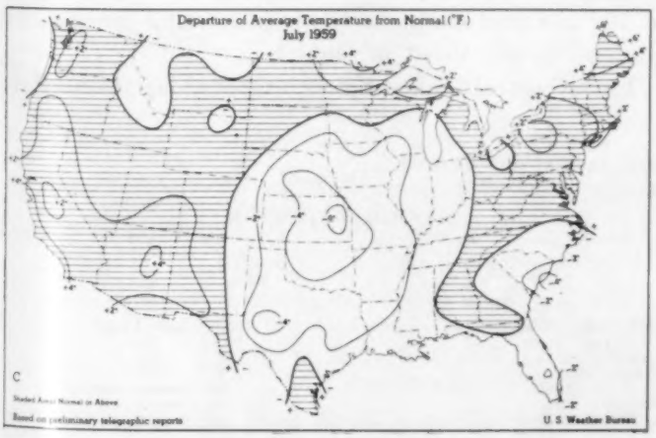
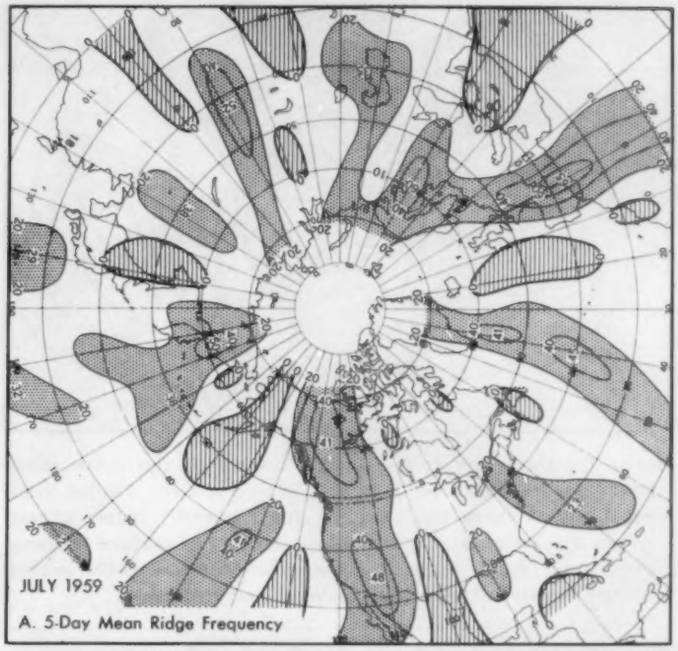


FIGURE 4.—Departure of average surface temperature from normal (°F.) for July 1959. Central States were cool, but record heat was observed in Arizona and California. (From [10] Aug. 3.)



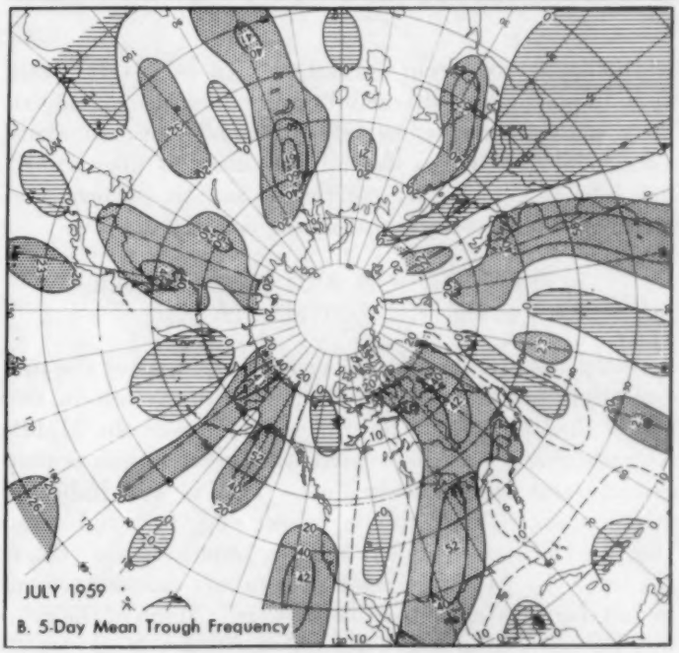


FIGURE 5.—Percent of the time that (A) ridges and (B) troughs on 5-day mean 700-mb. charts were located within 10° longitude intervals at latitudes from 20° N. to 70° N. for July 1959. The data were adjusted to an equivalent basis with 10° at 50° N. as the unit. Isoline interval is 20 percent; areas with frequency greater than 20 percent are stippled; zero areas are hatched. Note the high frequency of troughs in the central United States and ridges in the Far West.

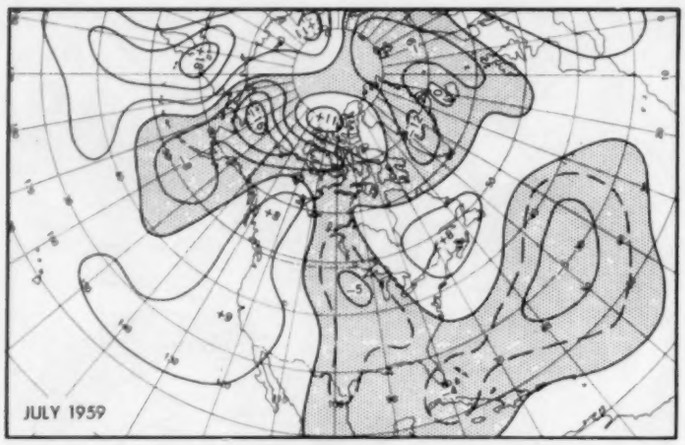


FIGURE 6.—Departure from monthly normal of the mean thickness (700–1000 mb.) for July 1959 in tens of feet. Isoline interval is 50 ft. with intermediate lines dashed. Below normal values are stippled. Pattern is similar to the surface temperature chart (fig. 4) with air masses generally cool in the central United States but warm in the West and Northeast.

height departures from normal, though relatively small, were important in determining the weather anomalies. They were oriented effectively to produce an anomalous, meridional flow across the normal thickness lines (1000–700 mb. [9], generally northerly in the western and southerly in the eastern half of the country.

3. PRECIPITATION AND TEMPERATURE OVER THE UNITED STATES

Except for the showers in Arizona, the bulk of the precipitation fell east of the 30-day mean trough in two rather distinct areas. Record amounts over the Middle Atlantic States were associated with greater than normal easterly and southeasterly flow. This is well indicated by the anomalies of 700-mb. height (fig. 1), wind speed (fig. 9), thickness (fig. 6), and sea level pressure (Chart XI in [8]). Moisture-laden tropical air masses predominated, facilitating convective showers. Much of the rainfall was associated with cold and stationary fronts which were located in this area more than half of the days during July (fig. 10). Very little rainfall was directly related to Lows, which were confined mainly to the higher latitudes (fig. 11B). However, one cyclone, hurricane Cindy, entered the South Carolina coast on July 8. Although it subsequently weakened, it produced copious rainfall along the Atlantic Seaboard from July 8 to 11. New England was north of the anomalous easterlies and was predominantly dry, except for the extreme southeastern part where near record precipitation amounts were reported. Weather, particularly visibility, was unusually bad in Nantucket, Mass., where fog or fog and rain occurred on 27 days of July.

The other major area of rainfall was centered over the Southern Plains and South-Central States. It was associated with the abnormally cyclonic 700-mb. flow (figs. 1, 5) and the anomalous southerly winds in the lower troposphere (fig. 1). No extratropical cyclones were observed in this area, and most rainfall resulted from heavy local shower activity and hurricane Debra. This short-lived, tropical cyclone deepened explosively on the 24th, just off the Texas coast (see Chart X in [8]) and filled rapidly as it moved northward into Oklahoma the next day. Debra produced heavy rainfall in Texas and Louisiana. Galveston, Tex., for example, reported over 5 inches.

West of the 30-day mean trough (fig. 1) and the region of frequent 5-day mean troughs (fig. 5) rainfall was generally light (fig. 3). A broad band of little or no rainfall extending from California northeastward to Montana was associated with persistent anticyclonic circulation and northerly anomalous flow at 700 mb. over the western United States (figs. 1, 5). Record (or near record) dryness was reported at many stations in Colorado, Idaho, Montana, the Dakotas, and Wyoming (table 1–B).

The light precipitation in the northern Appalachian Mountains was consistent with a weak mean ridge over that region (fig. 1), but an explanation for the pocket of light rainfall in the southern Appalachians is not readily extracted from the mean circulation. However, it does correspond with a center of anticyclonic relative vorticity at 700 mb. (not shown). Other interesting precipitation phenomena are listed in table 1.

In the central United States, except along the northern

TABLE 1.—*Abnormal precipitation (inches) for July 1959*

A. HEAVY AMOUNTS			B. LIGHT AMOUNTS		
Station	Monthly total	Remarks	Station	Monthly total	Remarks
Worcester, Mass.	8.11	Second wettest July.	Idaho Falls, Idaho	Trace	Record July.
Atlantic City, N.J.	15.69	Record month.	Great Falls, Mont.	0.04	Do.
Richmond, Va.	12.85	Second wettest July.	Havre, Mont.	.07	Do.
Greensboro, N.C.	9.81	Wettest July since 1927.	Helena, Mont.	.11	Do.
Winston-Salem, N.C.	10.86	Record July.	Kalispell, Mont.	.04	Second driest July.
Memphis, Tenn.	8.84	Do.	Sheridan, Wyo.	.08	Record July.
Oklahoma City, Okla.	8.44	Do.	Bismarck, N. Dak.	.41	Second driest July.
Tulsa, Okla.	9.85	Second wettest July.			
Port Arthur, Tex.	18.69	Do.			
Meridian, Miss.	15.29	Record July.			

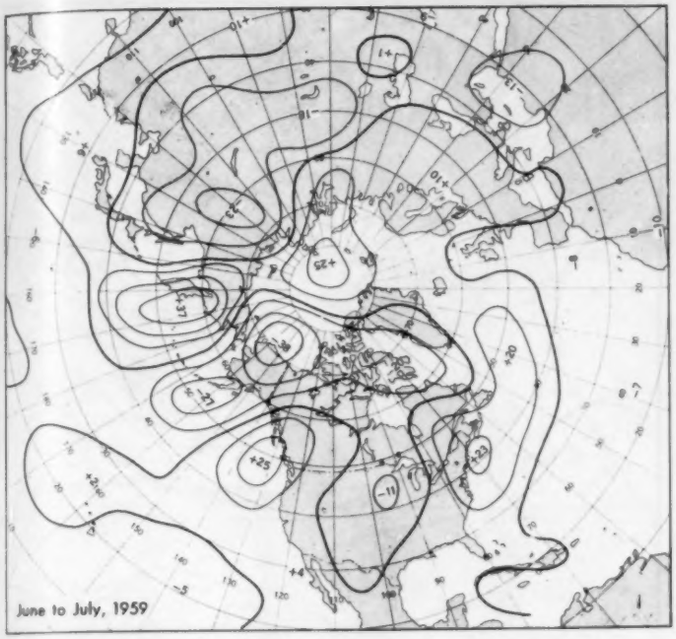


FIGURE 7.—Difference between 30-day mean 700-mb. height anomaly for June and July 1959 (July minus June) in tens of feet. Isoline interval is 100 feet. Over the United States the circulation departed from the common summertime pattern as heights fell (relative to normal) in the center and rose in the West and East.

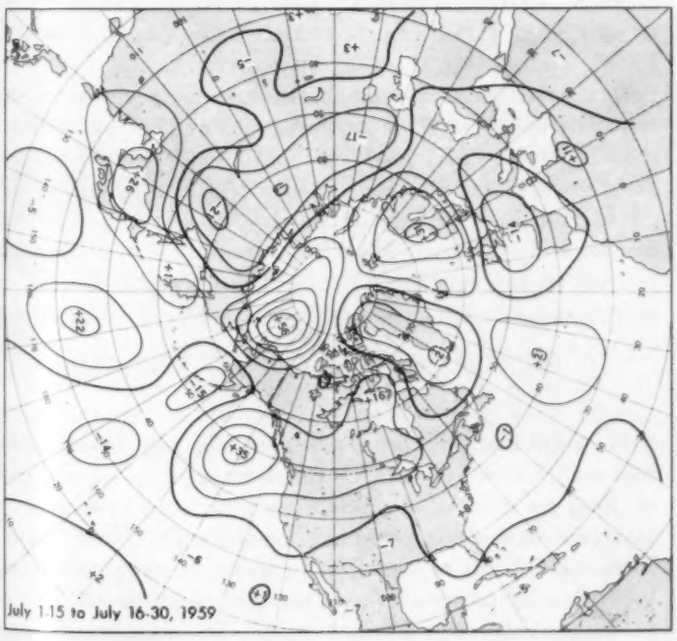
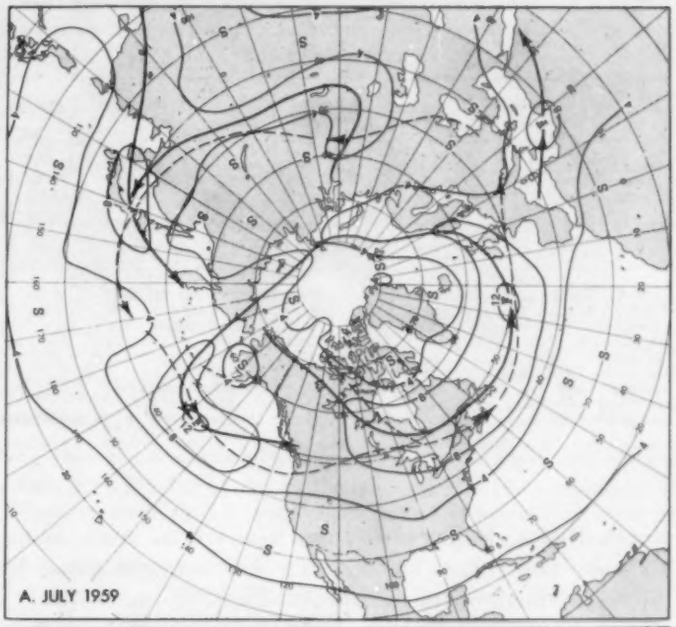


FIGURE 8.—Fifteen-day mean 700-mb. height change from July 1–15 to July 16–30, 1959 (July 16–30 minus July 1–15) in tens of feet. Isoline interval is 100 ft. Height changes were small over North America.



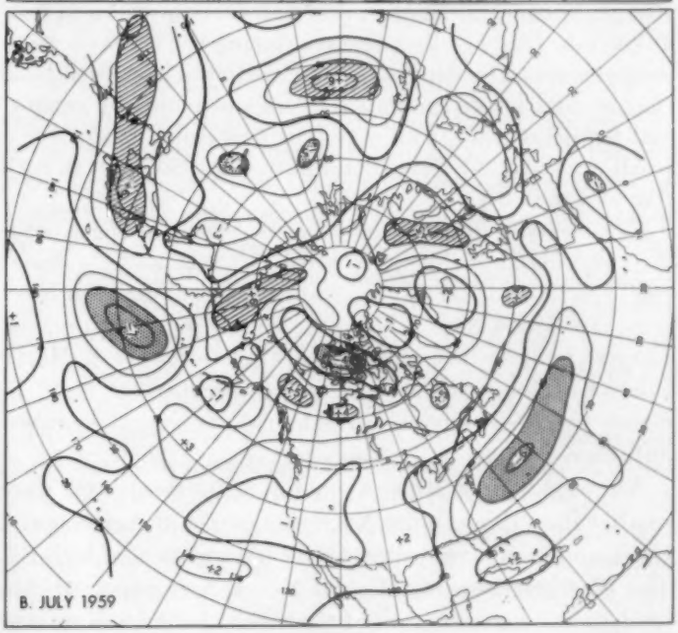


FIGURE 9.—(A) Mean 700-mb. isotachs and (B) departures from monthly normal wind speeds, both in meters per second, for July 1959. Solid arrows in (A) indicate principal axes of maximum winds, and dashed arrows give their normal July positions. Areas with anomalies greater than 4 m.p.s. are stippled. Wind speeds averaged below normal over eastern United States, reflecting the prevailing easterly regime in that area.

border, the weather was cooler than normal (fig. 4). In parts of Texas, Kansas, Nebraska, and Missouri, where temperatures averaged below normal every week [10], monthly anomalies were the largest and mean temperatures approached near record lows (table 2). This was marked cooling, relative to normal, from the previous month [4], and it was as much as four classes colder at

TABLE 2.—*Abnormal surface temperatures (°F.) for July 1959*

Station	Monthly average	Remarks
A. COOL WEATHER		
Midland, Tex.	78.1	Equalled July record.
Topeka, Kans.	74.4	Both cool and wet.
Lincoln, Nebr.	75.1	Do.
B. HOT WEATHER		
Caribou, Maine	69.5	Second warmest July.
Winslow, Ariz.	80.9	Record July.
Yuma, Ariz.	96.7	Record month.
Blue Canyon, Calif.	72.8	Do.
Los Angeles, Calif.	75.1	Do.
San Diego, Calif.	73.7	Record July.
Burbank, Calif.	77.2	Second warmest July.
Mt. Shasta, Calif.	70.8	Do.
Red Bluff, Calif.	85.3	Maximum 100° F. 24 consecutive days.

North Platte, Nebr. (from much above to much below normal) (fig. 12). This cooling was brought about by the replacement of the continental High of June by July's trough, subnormal heights, and northerly anomalous flow at 700 mb. (fig. 1). In this area during the summer mean temperatures are depressed by cloudiness and precipitation, so that heavy rainfall is usually attended by subnormal temperatures. This was true this July, except for a small area around Nebraska where the weather was both cool and dry.

Very warm temperatures were observed in the Northeast where 700-mb. heights were 140 feet above normal (fig. 1); 1000 to 700 mb. thicknesses (fig. 6) and sea level pressures (see Chart XI in [8]) were also above normal. Largest temperature departures were in northern Maine where Caribou, with an average temperature of 69.5° F. which is 5.5° F. above normal, had its second warmest July on record.

Although temperature anomalies in the Southwest were smaller than those in the Northeast, parts of Arizona and California had a record-breaking heat wave (table 2–B). The Weather Bureau station at Yuma, Ariz. reported that "July was a continuation of the record heat wave started in June. The monthly mean temperature of 96.7° F. made this not only the highest in July, but also the highest ever recorded for any month at Yuma. . . . For 30 consecutive days, from June 30th to July 29th inclusive, the minimum temperature was 81° or higher." Incidentally, the maximum temperatures for this July at Yuma averaged 109.4° F. Factors contributing to this heat wave were the weakness of the trough along the west coast and the locally above normal 700-mb. heights (fig. 1) and thicknesses (1000–700 mb.) (fig. 6). Of equal importance perhaps, was the adiabatic warming associated with subsidence underneath the upper-level High and with downslope northerly winds from the interior, as indicated by the 700-mb. chart (fig. 1) and anomalous thermal wind (fig. 6).

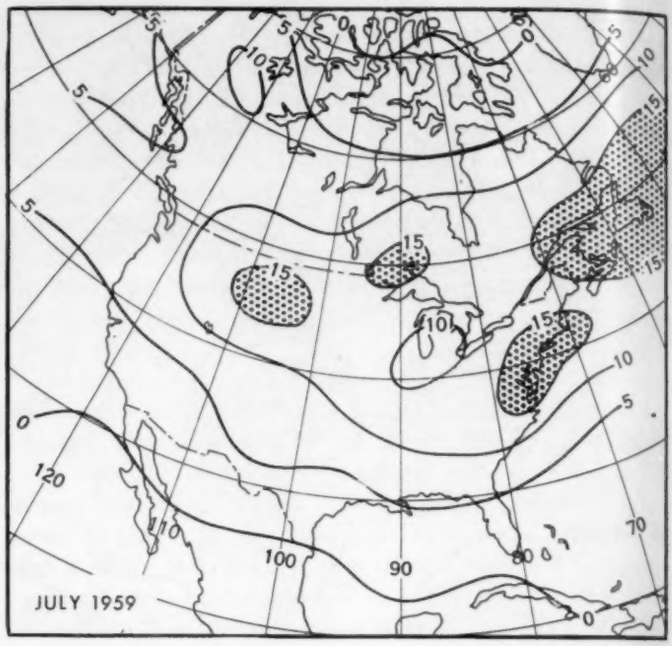


FIGURE 10.—Number of days in July 1959 with fronts of any type within unit squares (with sides approximately 500 miles). All frontal positions are taken from *Daily Weather Map*, 1:00 p.m. EST. Areas with 15 or more days with fronts are stippled. Active fronts were often located in the Middle Atlantic States.

4. CYCLONE AND ANTICYCLONE TRACKS

It has already been pointed out in section 3 that few cyclones affected the United States this July. There were some incipient Lows in the North Central States, but most cyclonic activity was in Canada, north of the mean jet stream (fig. 11B). Lows were common in the central Pacific and eastern Canadian mean troughs.

The principal anticyclone track this July was across central United States somewhat south of (but parallel to) the climatologically preferred path [5]. This deviation from normal was also associated with the unusual 700-mb. circulation and furthered the dry, cool weather in Nebraska and vicinity. Most high pressure centers first appeared in the northern Rocky Mountains as masses of maritime polar air separated from the Pacific anticyclone.

5. POSSIBLE CAUSES OF THE ABNORMAL CIRCULATION

Why did a circulation not climatologically favored prevail over the western Northern Hemisphere for a period as long as a month? What factors were operating to maintain this rarely observed circulation? Mountain effects are difficult to assess, but various heating terms may be estimated using methods similar to those of Wexler [11] and Aubert and Winston [1]. Diabatic heating may be inferred (but with some doubt) from figure 13 which has the 30-day mean thickness lines (700–1000 mb.) superimposed on the mean 700-mb. contours. Where the indicated advection of thickness was large (contours and

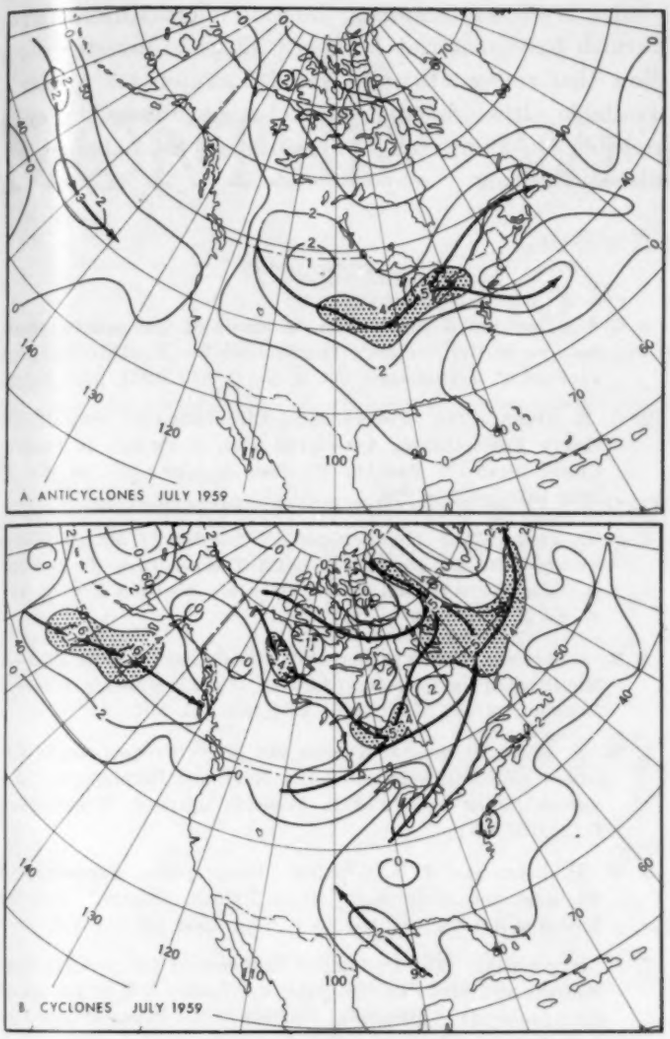


FIGURE 11.—Number of (A) anticyclone passages and (B) cyclone passages (within equal area quadrilaterals of 66,000 n.mi.²) during July 1959. Primary tracks are indicated by solid arrows. Isoline interval is 2. Cyclones were generally confined to Canada, but most of the anticyclones had trajectories south of normal over central United States.

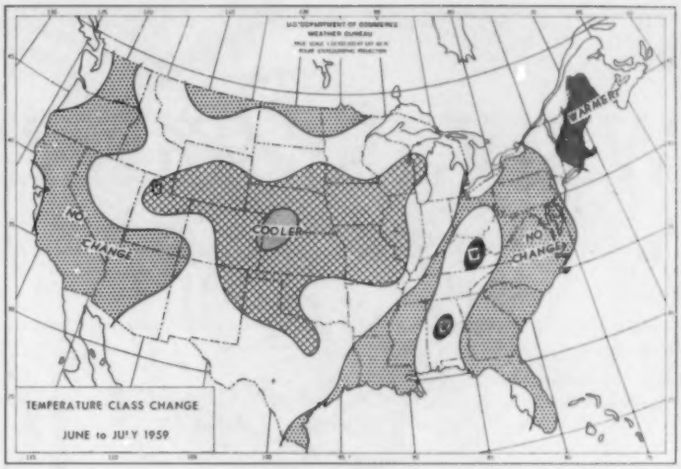


FIGURE 12.—Number of classes the anomaly of surface temperature changed from June to July 1959. Areas where the anomaly in July was 2 or 3 classes warmer are hatched; 2 or 3 classes cooler, cross-hatched; 4 classes cooler, fine stippled; and areas of no change, coarse stippled. July was cooler than June in the Midwest, relative to normal.

thickness lines out of phase), there may have been present any one of the following three processes: diabatic heating and cooling, adiabatic warming and cooling through vertical motion, and diffusion of heat by the daily eddies. These effects are difficult to isolate, in this instance, but an attempt will be made to deduce the most important one.

The two main regions of indicated advection were over northwestern United States and the western Pacific south of the Kamchatka Peninsula. We are forced to ignore the eddy terms since they are unknown, even though they may be quite large in some areas. As in a previous study [3], the monthly mean vertical motion was approximated by averaging 30 daily values at 600 mb. computed from initial data from the Joint Numerical Weather Prediction Unit's baroclinic model. The vertical motion was found to be small, not greater than ±1 m./sec. in the two areas just referred to. It should be pointed out that this vertical motion was computed by the adiabatic assumption and perhaps cannot be used to assess heating terms. The cold advection indicated in northwestern United States suggests diabatic heating (fig. 13A). The same is true in the long-period normal (fig. 13B). Aubert and Winston [1] found that western United States was an area where the atmosphere gains heat in July.

In the western Pacific the indicated warm advection suggests diabatic cooling. (This is also the normal situation, but visual comparison of the two charts in fig. 13 reveals that the indicated advection was much stronger than normal this July.) Air ascending at the average rate of 1 mm./sec. and cooling at the dry adiabatic rate would cool 26° C./30 days. The apparent increase in temperature due to horizontal advection in the area is calculated of the order of 100° C./30 days. The net calculated temperature change due to warming by advection and cooling by vertical motion exceeds any possible local temperature change and suggests that diabatic cooling occurred. This cooling would produce anticyclogenesis at sea level downstream from the center of the heat sink [7] and, at least partially, explain the abnormally high pressure observed in the western Bering Sea. A positive sea level pressure anomaly (approximately 8 mb.), quite similar to the 700-mb. height anomaly (fig. 1), was centered at 52° N., 170° E.

Severe criticism may be leveled at this hypothesis because of the drastic assumptions made and the absence of more quantitative work. However, this section was in-

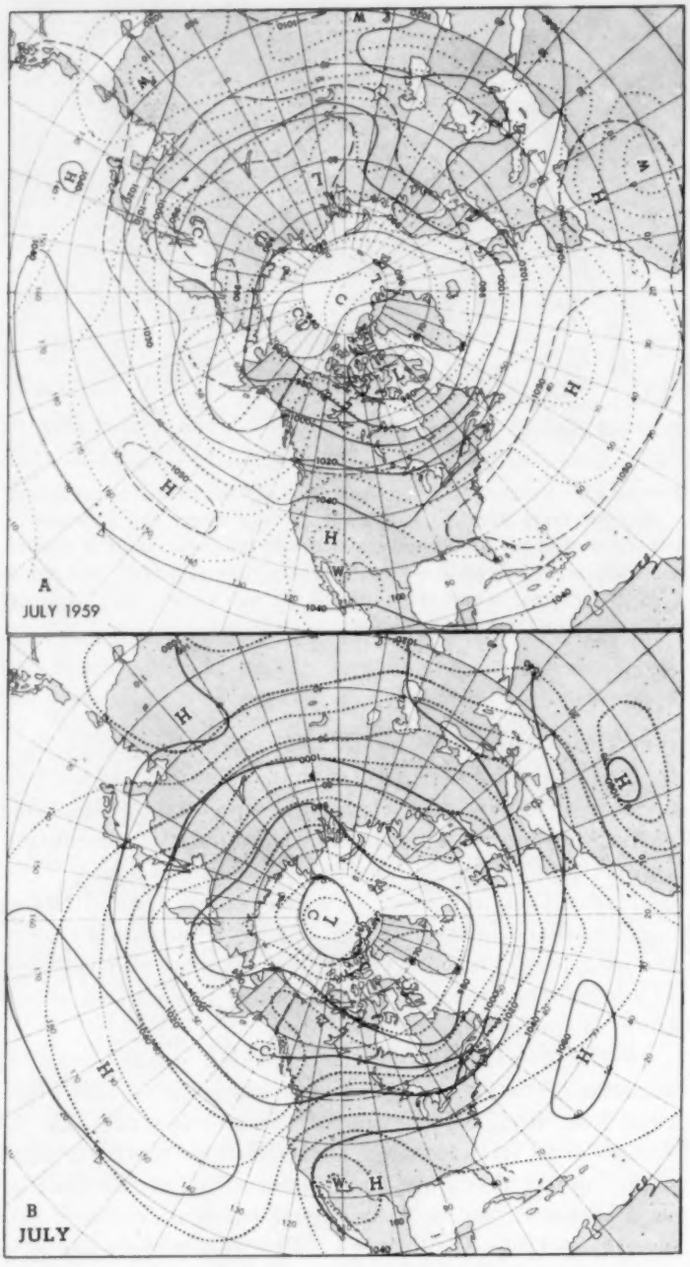


FIGURE 13.—Mean 700-mb. contours (solid) and thickness (1000–700 mb.) (dotted), both in tens of feet. (A) is for July 1959 and (B) is the July normal chart. 700-mb. contour interval is 200 ft.; thickness isoline interval is 100 ft. This July there was strong indicated advection of thickness (mean virtual temperature) in the western Pacific.

cluded to focus attention on the observed conditions and to furnish basic material for more complete investigations. Now that computations of vertical motion are routinely available, although based on adiabatic processes, it may be possible to obtain rough quantitative estimates of the diabatic heating.

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U.S. GOVERNMENT PRINTING OFFICE: 1959